SIMULATION OF AQUIFER TESTS AND GROUND-WATER FLOWPATHS AT THE LOCAL SCALE IN FRACTURED SHALES AND SANDSTONES OF THE BRUNSWICK GROUP AND LOCKATONG FORMATION, LANSDALE, MONTGOMERY COUNTY, PENNSYLVANIA

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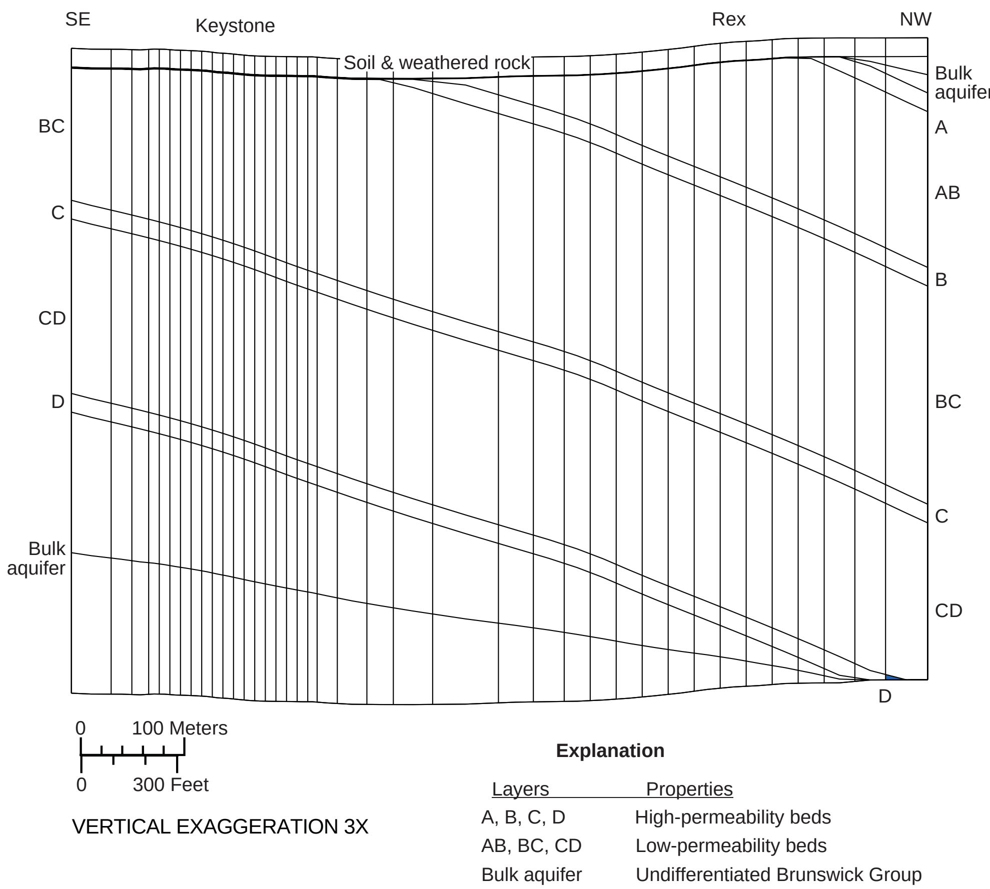

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Cover: Cross-section of a three-dimensional ground-water flow model showing dipping high- and low-permeability beds of the Brunswick Group in northwestern Lansdale, Pennsylvania (see figure 24 of this report, and related discussion). “Keystone” and “Rex” designate properties where aquifer tests were conducted to provide drawdown and recovery data for calibration of the model.

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Lemoyne, Pennsylvania
2000

U.S. DEPARTMENT OF THE INTERIOR

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ILLUSTRATIONS

ILLUSTRATIONS--continued

ILLUSTRATIONS--continued

TABLES

CONVERSION FACTORS, ABBREVIATIONS, AND VERTICAL DATUM

Multiply	By	To obtain
	Length	
inch (in)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square foot (ft^2)	0.09290	square meter
square mile (mi^2)	2.590	square kilometer
	Volume	
gallon (gal)	3.785	liter
gallon (gal)	0.003785	cubic meter
	Flow rate	
foot per day (ft/d)	0.3048	meter per day
gallon per minute (gal/min)	0.06309	liter per second
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day
	Transmissivity	
foot squared per day (ft^2/d)	0.09290	meter squared per day

Vertical datum: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929—a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

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ABSTRACT

The U.S. Geological Survey, as part of technical assistance to the U.S. Environmental Protection Agency, has constructed and calibrated models of local-scale ground-water flow in and near Lansdale, Pa., where numerous sources of industrial contamination have been consolidated into the North Penn Area 6 Superfund Site. The local-scale models incorporate hydrogeologic structure of northwest-dipping beds with uniform hydraulic properties identified in previous studies. Computations associated with mapping the dipping-bed structure into the three-dimensional model grid are handled by a preprocessor using a programmed geographic information system (GIS). Hydraulic properties are identified by calibration of the models using measured water levels during pumping and recovery from aquifer tests at three sites. Reduced flow across low-permeability beds is explicitly simulated. The dipping high-permeability beds are extensive in the strike direction but are of limited extent in the dip direction. This model structure yields ground-water-flow patterns characteristic of anisotropic aquifers; preferred flow is in the strike direction. The transmissivities of high-permeability beds in the local-scale models range from 142 to 1,900 ft^2/d (feet squared per day) (13 to 177 m^2/d). The hydraulic conductivities of low-permeability parts of the aquifer range from 9.6×10^{-4} to 0.26 ft / d (feet per day) (2.9×10^{-4} to 0.079 m / d). Storage coefficients and specific storage are very low, indicating the confined nature of the aquifer system. The calibrated models are used to simulate contributing areas of wells under alternative, hypothetical ground-water-management practices. Predictive contributing areas indicate the general characteristics of ground-water flow towards wells in the Lansdale area. Recharge to wells in Lansdale generally comes from infiltration near the well and over an area that extends upgradient from the well. The contributing areas for two wells pumping at 10 gal / min (gallons per minute) extend about 1,500 ft (feet) upgradient from the wells. The contributing area is more complex at ground-water divides and can extend in more than one direction to capture recharge from more than 3,300 ft away, for pumping at a rate of 30 gal / min. Locally, all recharge in the area of the pumping well is not captured; recharge in the downgradient direction about 150 ft from the pumping well will flow to other discharge locations.

INTRODUCTION

Ground water in the area of the Borough of Lansdale, Pa., has been withdrawn since the early 20th century for use as drinking water and for industrial supply. In 1979, water from public-supply wells in the area was found to be contaminated with trichloroethylene (TCE), tetrachloroethylene (PCE), and other human-made organic compounds (CH2M Hill, 1991). Through additional sampling, an area of ground-water contamination was identified, and the site, known as North Penn Area 6, was placed on the National Priority List (NPL) by the U.S. Environmental Protection Agency (USEPA). The North Penn Area 6 site encompasses about 3 mi^2 (square miles) [2.6 km^2 (square kilometers)] and includes at least six sources of contamination on separately-owned properties largely within the Borough of Lansdale (CH2M Hill, 1991). The site is located on the U.S. Geological Survey (USGS) Lansdale and Telford 7.5-minute topographic quadrangle maps (fig. 1).

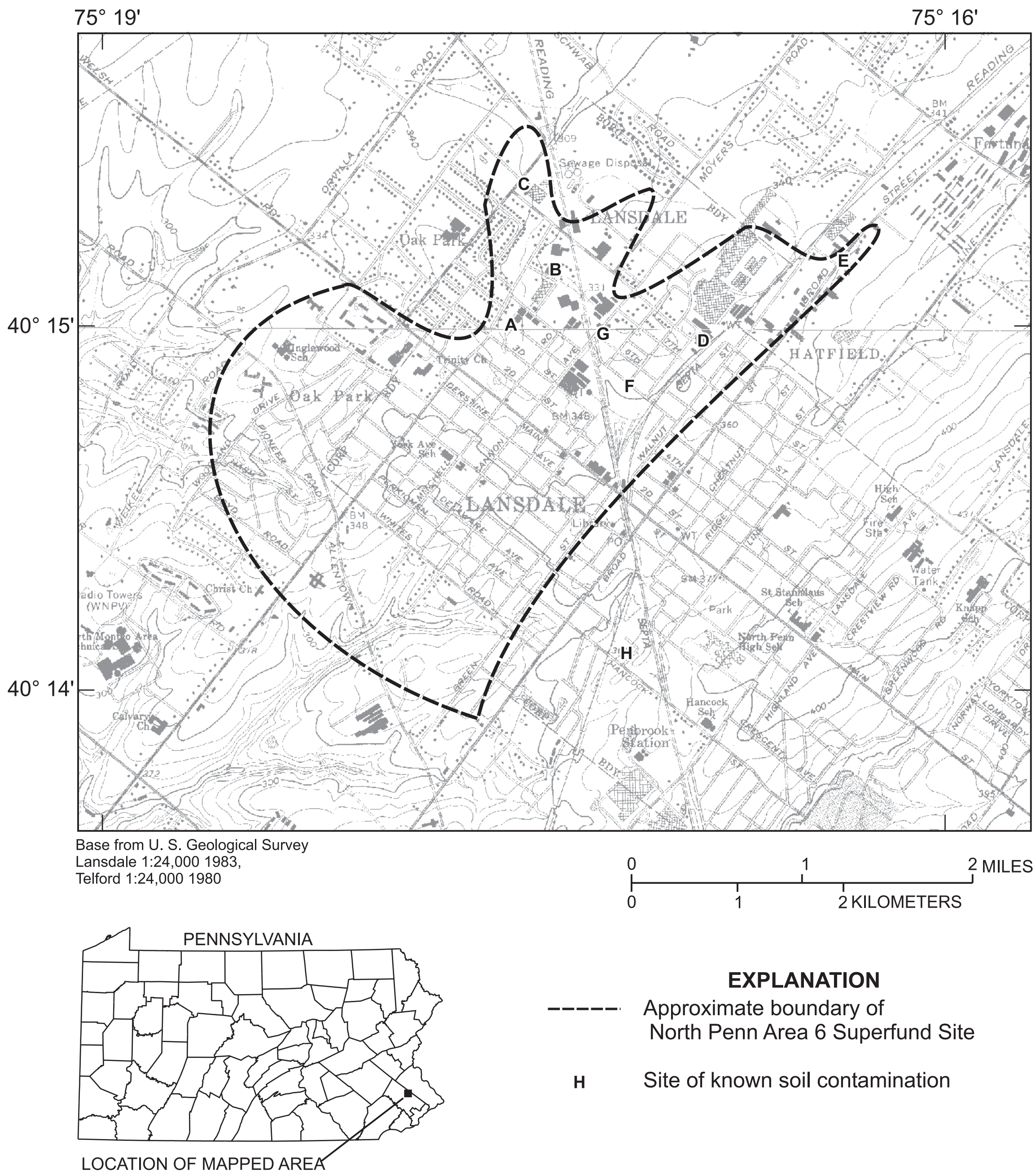


Figure 1.-- Location of North Penn Area 6 site, Lansdale, Pa.

Since 1995, abandonment of public-supply wells in favor of an alternative surface-water supply source and closure of industrial facilities has changed the location and rate of ground-water withdrawals in Lansdale. Concerned about contaminant migration, the USEPA needed information about the effects of these changes in water use on the direction of ground-water flow. In 1995, the USGS, in cooperation with USEPA, began a study to describe the ground-water system and simulate ground-water flow on a regional scale using a numerical model in the area of Lansdale. Data collected for the study from 1996 through 1998 included geophysical logs of wells, water levels in wells, streamflow measurements, aquifer-interval-isolation tests, and multiple-well aquifer tests (Conger, 1999; Senior and Goode, 1999). This work was done to assist the USEPA in preparing a remedial investigation and feasibility study (RI/FS) of the North Penn Area 6 site (Black & Veatch Waste Science, Inc., 1994, 1998).

The numerical model used by USGS (Senior and Goode, 1999) to simulate ground-water flow in the area of Lansdale provided estimates of bulk aquifer transmissivity and general ground-water-flow paths on a regional scale, but not at the local or site scale. The regional-scale model structure did not incorporate local heterogeneity or the geologic structure of dipping beds, aquifer characteristics that appear to affect local ground-water flow as determined from aquifer tests at four properties in North Penn Area 6 in 1997 (Senior and Goode, 1999). Therefore, the USGS proposed, in late 1999, additional simulations to more accurately simulate ground-water flow at the local scale in selected areas where pumping may be used as part of the ground-water remediation.

Purpose and Scope

This report presents numerical simulations of ground-water flow using the porous-media model MODFLOW (Harbaugh and McDonald, 1996) at the local scale for two areas in and near Lansdale, Pa. The simulations are based on a model structure that includes the geologic structure of dipping beds.The automatic, nonlinear optimization program, MODFLOWP (Hill, 1992), is used to calibrate the model to water levels measured during aquifer tests done in 1997. Contributing areas for wells pumped during these aquifer tests and drawdown in the pumped well and observation wells are simulated.

Previous Work

Work done by USGS for USEPA on North Penn Area 6, Lansdale, Pa., is described in reports by Goode and Senior (1998), Conger (1999), and Senior and Goode (1999). Senior and Goode (1999) discuss results of numerous previous studies in the Lansdale area. Aquifer tests at the J.W. Rex Co. property are summarized in the report prepared by QST, Inc. (1998).

Acknowledgments

The additional work done for simulation of ground-water flow at the local scale was initiated from reviews of the regional scale model by Gregory Ham and Kathy Davies of the USEPA and Lusheng Yan of Black & Veatch Waste Science, Inc. Data for aquifer tests at the J.W. Rex Co. property were provided by Brian Loughnane of ESE, Inc.

HYDROGEOLOGIC SETTING

The study area in and near Lansdale is in the Gettysburg-Newark Lowland Section of the Piedmont Physiographic Province. The North Penn Area 6 site and surrounding area are underlain by sedimentary rocks of the Lockatong Formation and lower beds of the Brunswick Group of the Newark Supergroup (Lyttle and Epstein, 1987) (fig. 2). Sediments of the Newark Supergroup were deposited in a rift basin during the Triassic age (260 million years ago). Following deposition, sediments in the Newark Basin were buried, compacted, and faulted. Commonly, the Lockatong Formation is relatively resistant to erosion and tends to form ridges that rise above flat or rolling topography underlain by rocks of the Brunswick Group. Lansdale and the North Penn Area 6 site are underlain mostly by rocks of the Brunswick Group and are on relatively flat upland terrain that is a surface-water divide between Wissahickon Creek to the south, Towamencin Creek to the west, and tributaries to the West Branch Neshaminy Creek to the north and northeast (figs. 1 and 2).

Geology

The Lockatong Formation consists of detrital sequences (cycles) of gray to black calcareous shale and siltstone, with some pyrite, and chemical sequences (cycles) of gray to black dolomitic siltstone and marlstone with lenses of pyritic limestone, overlain by massive gray to red siltstone with analcime (Lyttle and Epstein, 1987). Interbeds of reddish-brown, sandy siltstone have been mapped in the Lockatong Formation south of Lansdale (Lyttle and Epstein, 1987). The Lockatong Formation overlies the Stockton Formation, which consists of gray to reddish-brown sandstones, shales, and siltstones. Contacts between the Lockatong Formation and the overlying Brunswick Group are conformable and gradational, and the two formations may interfinger (Lyttle and Epstein, 1987). The lower beds of the Brunswick Group consist predominantly of homogeneous, soft, red to reddish-brown and gray to greenish-gray mudstones and clay- and mud-shales, with some fine-grained sandstones and siltstones. Some beds are micaceous. Bedding is irregular and wavy. Interbedded silt-shales and siltstones are moderately well sorted. The Brunswick Group rocks contain detrital cycles of medium- to dark-gray and olive- to greenish-gray, thin-bedded and evenly bedded shale and siltstone, similar to the underlying Lockatong Formation. Red-brown shale, red-brown sandstone, and gray shale are the most frequently reported rock types in drillers' logs for monitor wells drilled in 1997 in Lansdale (Black & Veatch Waste Science, Inc., written commun., 1997).

Bedding in the Newark Basin generally strikes northeast and dips to the northwest. The regional homoclinal dip has been cut by normal and strike-slip faults and warped by transverse folds (Schlische, 1992). Many faults with small displacements have not been mapped. The beds of the Brunswick Group and Lockatong Formation generally strike northeast and dip shallowly to the northwest in the vicinity of the North Penn Area 6 site; strike gradually shifts from northeast in central Lansdale to east-northeast in the area south of Lansdale near Upper Gwynedd Township (fig. 2) (Longwill and Wood, 1965). Thin shale marker beds in the Brunswick Group identified by elevated natural-gamma activity on geophysical logs can be correlated over distances of 1,000 ft [300 m (meters)] or more. High natural-gamma activity typically is associated with thin gray shale beds. Correlation of natural-gamma activity in logs collected by USGS in and near Lansdale shows that these shale beds strike 48^{o} to 60^{o} northeast and dip 6^{o} to 30^{o} northwest; the average dip is about 11^{o} (Conger, 1999).

Ground-Water System

Ground water in the rocks underlying Lansdale and the North Penn Area 6 site originates from infiltration of precipitation. After infiltrating through soil and saprolite (extensively weathered rock), ground water moves through fractures and openings in the shale and siltstone bedrock. Depth to bedrock is commonly less than 20 ft (6 m) below land surface. The soil, saprolite, and individual beds of the sedimentary bedrock form a layered aquifer. The degree of hydraulic connection between the layers varies. Hydraulic properties of the soil, saprolite, and individual beds of the underlying sedimentary bedrock differ. Primary porosity, permeability, and storage in the Triassic-age sedimentary bedrock is very low.

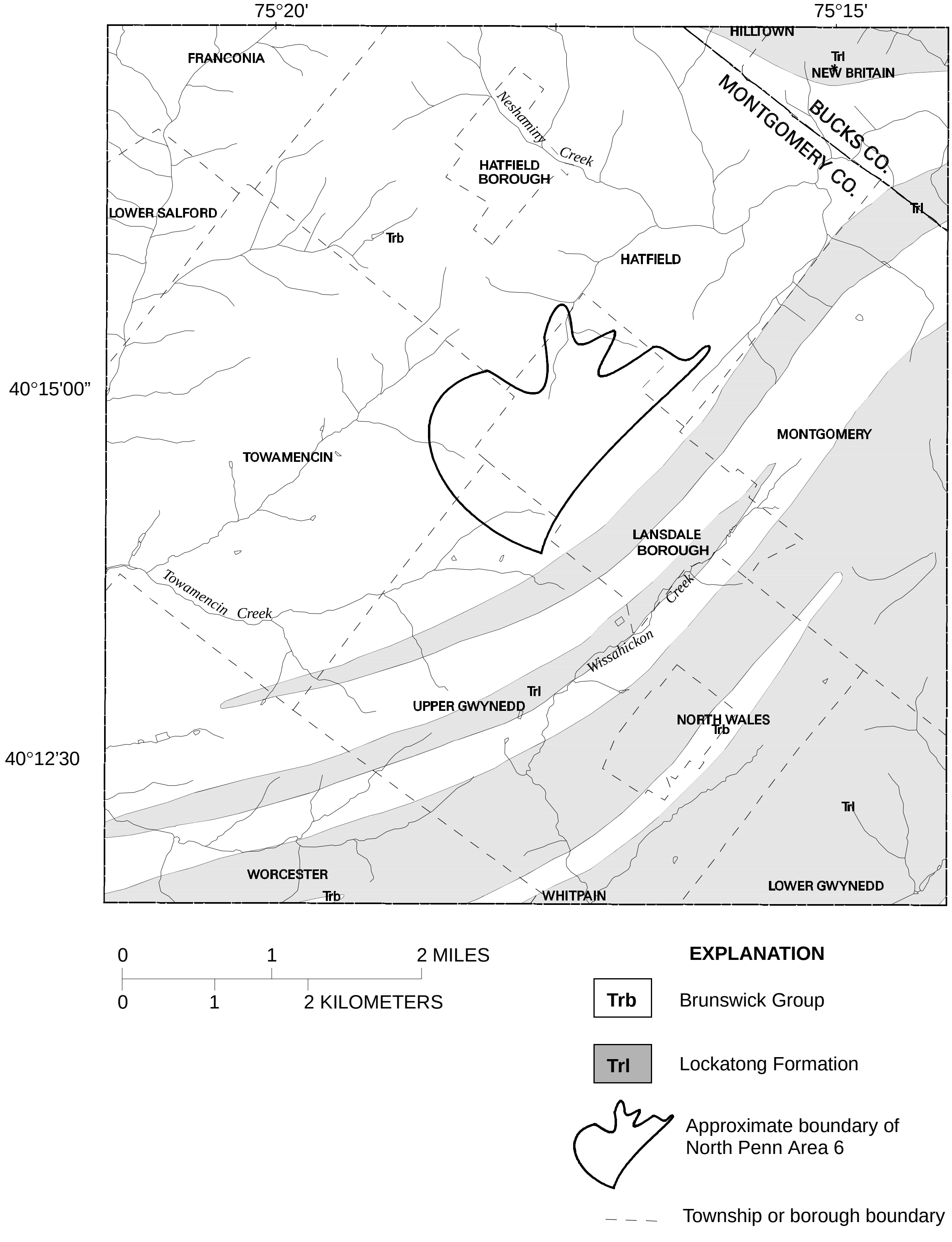


Figure 2.-- Bedrock geology in area of Lansdale, Pa.

Recharge to areas underlain by shales, siltstones, and sandstones of the Newark Basin tends to be lower than recharge to other areas of the Piedmont in southeastern Pennsylvania. Reported estimates of recharge to areas underlain by the Triassic sedimentary rocks of the Newark Basin range from 6 to 12 in. (153 to 305 mm) (Sloto and Schreffler, 1994). Senior and Goode (1999) estimated a recharge rate of 8.3 in/yr (212 mm/yr) in the Lansdale area by calibration of a regional ground-water-flow model using measured hydraulic heads and streamflow. The permeability of soils, saprolite, and underlying bedrock of the Triassic sedimentary rocks of the Newark Basin probably is lower than in areas underlain by other rocks in the Piedmont.

Shallow and deep ground-water-flow systems may be present at the site. Water from the shallow system likely discharges locally to streams and leaks downward to the deep ground-water-flow system. Shallow and deep ground water generally flows in a direction similar to the topographic gradient. Deep ground water discharges to streams and to pumping wells; the natural direction of shallow and deep ground-water flow is altered by pumping. Pumping from deep zones may induce downward flow from shallow zones. Water in the shallowest part of the sedimentary-rock aquifer may be under unconfined (water-table) or partially confined conditions; the unconfined part of the aquifer probably is thin and is difficult to delineate. In some areas, perched water is present at shallow depths [less than 50 ft (15 m)]. Water in the deeper part of the aquifer generally is confined or partially confined, resulting in artesian conditions.

The conceptual model of the ground-water system in the study area consists of dipping, layered fractured rocks with ground-water flow within partings developed primarily along bedding planes. Vertical fractures generally do not cut extensively across beds but may provide local routes of ground-water flow or leakage between beds (fig. 3).

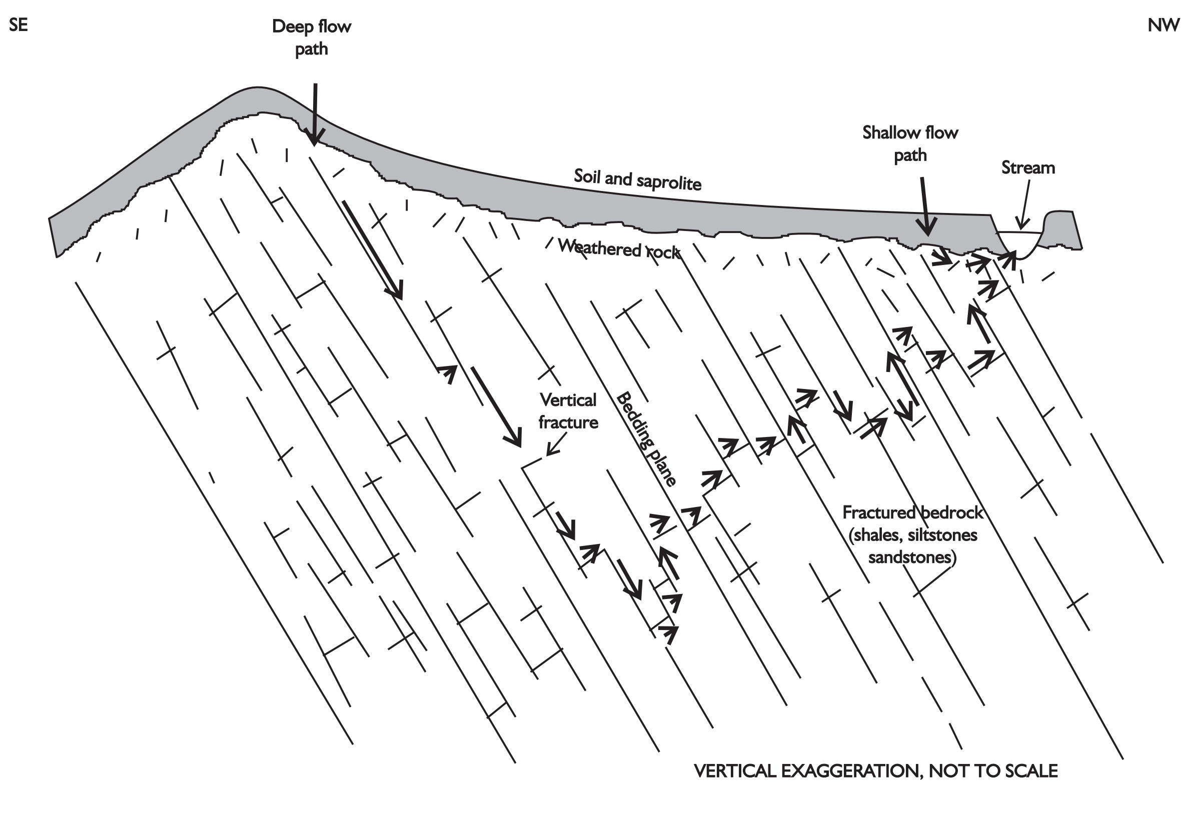


Figure 3.-- Schematic of conceptual model of ground-water-flow system for fractured sedimentary rock aquifer with dipping beds.

Regional ground-water flow in the area of Lansdale, Pa., was simulated using the porous-media model MODFLOW and calibrated using MODFLOWP for steady-state conditions (Senior and Goode, 1999). All recharge entering the top surface of the model area was assumed to discharge to wells or streams. Streams and topographic divides were assumed to form no-flow boundary conditions for the model (fig. 4). The model was configured with three vertical layers and two horizontal zones and a horizontal grid with uniform cell size of 328 ft x 328 ft (100 m x 100 m). The top layer of the 3-layer model represented weathered rock and soil, was 40-ft (13-m) thick, and was assumed to be isotropic and homogeneous. The bottom two layers of the model represented unweathered fractured rock, were each 328-ft (100-m) thick, and were assumed to be anisotropic with highest hydraulic conductivity (K) in the strike direction. Two zones of different hydraulic conductivity (K) were used for parts of the aquifer dominated by rocks of the Brunswick Group, in one zone, and the Lockatong Formation in the other. Senior and Goode (1999) assumed the regional-scale ground-water flow in the fractures could be modeled approximately using the porous-media model. In this model, the K values used represent the regional-scale effective properties of the bulk aquifer. Local-scale flow dominated by high-K zones within the aquifer could not be simulated in detail with this model.

Calibration of the regional model to 1996 conditions yielded estimates of the regional-scale K of the formations (table 1; Senior and Goode, 1999). The transmissivity of the weathered zone (layer 1) was estimated as 0.16 ft / d (hydraulic conductivity) x 40 ft (layer thickness) = 6.4 ft^2 / d (0.59 m^2 / d). The transmissivity of the underlying Brunswick Group (layers 2 and 3) in the strike direction was estimated as 5.35 ft / d x 656 ft = 3,510 ft^2 / d (326 m^2 / d). The transmissivity of the Brunswick Group in the dip direction was estimated as 3,510 ft^2 / d x 0.090 = 316 ft^2 / d (29 m^2 / d). The geometric mean (square root of the product) of the directional transmissivities corresponds to the “effective” isotropic transmissivity controlling drawdown due to pumping (Kruseman and de Ridder, 1990, p. 134). For the Brunswick Group, the geometric mean transmissivity was about 1,050 ft^2 / d (97 m^2 / d). The transmissivity of the unweathered part of the Lockatong Formation (in layers 2 and 3) was similarly estimated as 732 ft^2 / d (68 m^2 / d) in the strike direction and 64 ft^2 / d (6 m^2 / d) in the dip direction; the geometric mean was 215 ft^2 / d (20 m^2 / d). Most of the water moving horizontally through the model did so in layers 2 and 3, which represent unweathered fractured rock. The transmissivity of the zone representing the Brunswick Group was higher than that of the Lockatong Formation zone.

Table 1.-- Optimum and approximate, individual, 95-percent confidence-interval values for hydraulic conductivity, anisotropy ratio, and recharge for calibrated simulation of ground-water flow in and near Lansdale, Pa.

[K, hydraulic conductivity; ft / d, feet per day; -, dimensionless; in. / yr, inches per year]

Parameter	(units)	Optimum value	Approximate, individual, 95-percent confidence interval: lower value	upper value
K - Brunswick Group[1]	(ft/d)	[2]5.35	4.04	7.05
K - Lockatong Formation[1]	(ft/d)	[2]1.12	0.89	1.40
K - weathered zone[3]	(ft/d)	0.16	0.01	2.00
Anisotropy ratio of bedrock[4]	-	0.090	0.060	0.119
Recharge	(in./yr)	8.3	7.9	8.8

[1] Model layers 2 and 3 representing unweathered bedrock.
[2] In strike direction of model layers 2 and 3.
[3] Model layer 1 representing soil and saprolite.
[4] K(across strike)/K(along strike) for model layers 2 and 3.

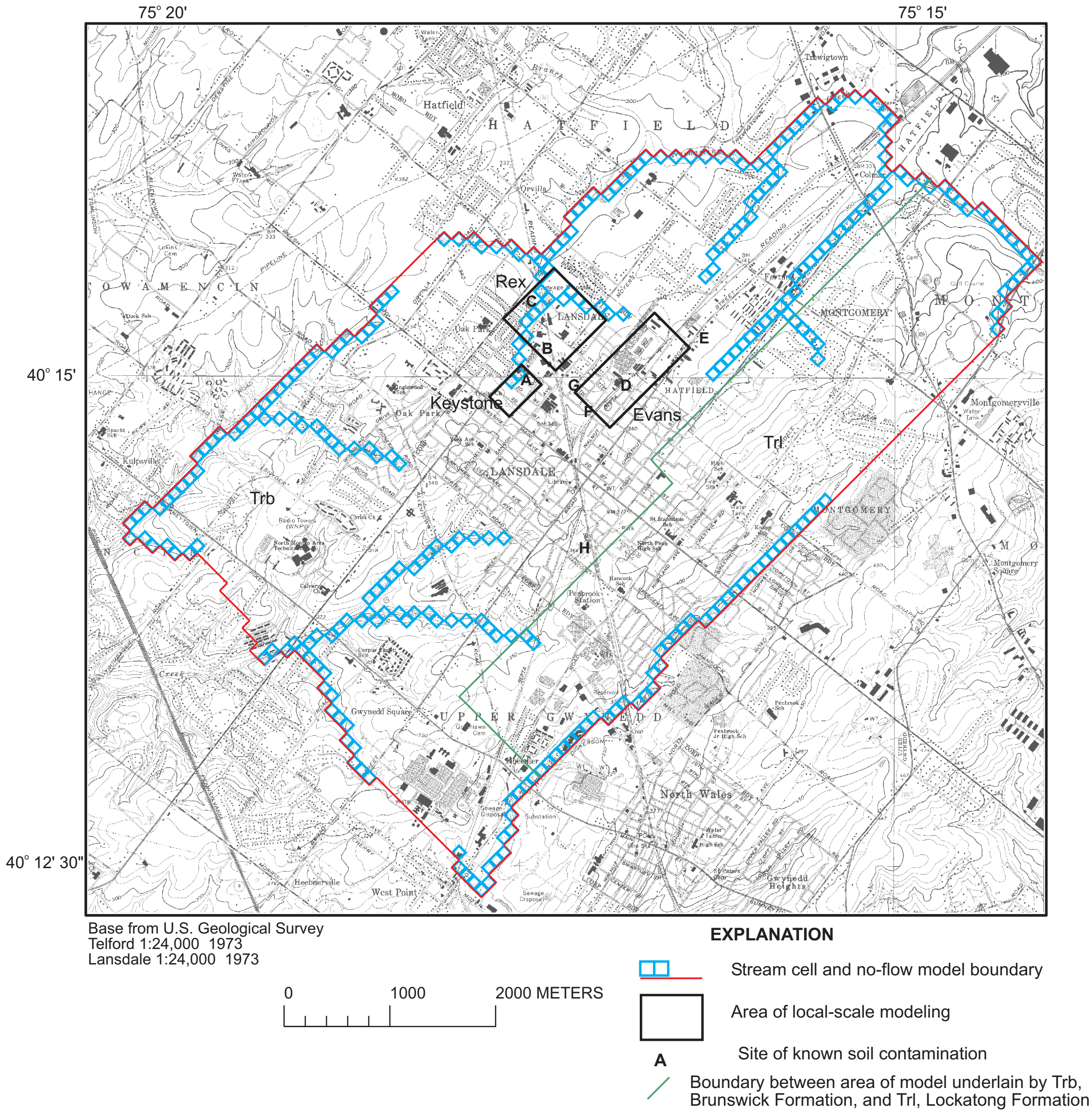


Figure 4.-- Boundaries and stream cells of regional-scale ground-water-flow model grid and areas of local-scale models at the John Evans and Sons property (Evans) in north-central Lansdale and at the Keystone Hydraulics property (Keystone) and the J.W. Rex Co. property (Rex) in northwestern Lansdale, Pa. Also shown are selected areas of soil contamination and the parts of the model area underlain by the Lockatong Formation (Trl) and Brunswick Group (Trb) (modified from Senior and Goode, 1999).

The calibrated flow model was used to simulate ground-water flow and hydraulic heads under three scenarios with different pumping conditions. It was assumed that the flow in a semi-confined aquifer system responds relatively quickly to changes in pumping rates; hence, a steady-state model was used. The first scenario, with no pumping, represents unstressed ground-water conditions with all recharge to the saturated zone discharging to streams as base flow. The second scenario, year 1994, represents periods with high pumping rates in the Lansdale area. The third scenario, year 1997, represents periods with moderate to high pumping rates that are less than those in the 1994 scenario, particularly for wells in the borough of Lansdale. Between 1994 and 1997, several public-supply wells were removed from service because a surface-water supply became available to North Penn Water Authority (NPWA), and several industrial wells were shut down due to plant closure. Overall pumping was a smaller fraction of average recharge (average ground-water discharge) in 1997 than in 1994 (Senior and Goode, 1999). Average recharge estimated for the 1996 calibration period was 8.3 in./yr (inches per year).

Particle tracking using MODPATH (Pollock, 1994) illustrates the paths of ground-water flow simulated by the regional flow model (fig. 5, Senior and Goode, 1999). On the basis of calibrated anisotropic transmissivities and vertical hydraulic conductivity and the computed three-dimensional hydraulic gradients, water particles were tracked through the flow system from recharge to discharge locations in streams or wells from eight areas of known soil contamination in and near Lansdale (table 2). The particle paths are skewed toward the direction of highest transmissivity (strike of the sedimentary beds), reflecting a regional anisotropy ratio of about 11 to 1.Hydraulic head in model layer 2 and recharge contributing areas for the 1997 simulation are shown in figure 5 (Senior and Goode, 1999).

Table 2.-- Selected sites and primary volatile organic compounds where soil contamination or probable sources of ground-water contamination have been identified in Lansdale, Pa. (Source of data: Black & Veatch Waste Science, Inc. 1994; Greg Ham, U.S. Environmental Protection Agency, written commun., 1997)

[TCE, trichloroethylene; PCE, tetrachloroethylene; VC, vinyl chloride]

Site code	Site name	Primary volatile organic compound(s) on site
A	Keystone Hydraulics	TCE,PCE,VC
B	Westside Industries	TCE,VC
C	J.W. Rex Co.	TCE,PCE,VC
D	John Evans and Sons	TCE,PCE
E	Royal Cleaners	PCE
F	Electra Products	PCE
G	Precision Rebuilding	TCE
H	Rogers Mechanical[1]	TCE

[1] Formerly the Tate Andale property.

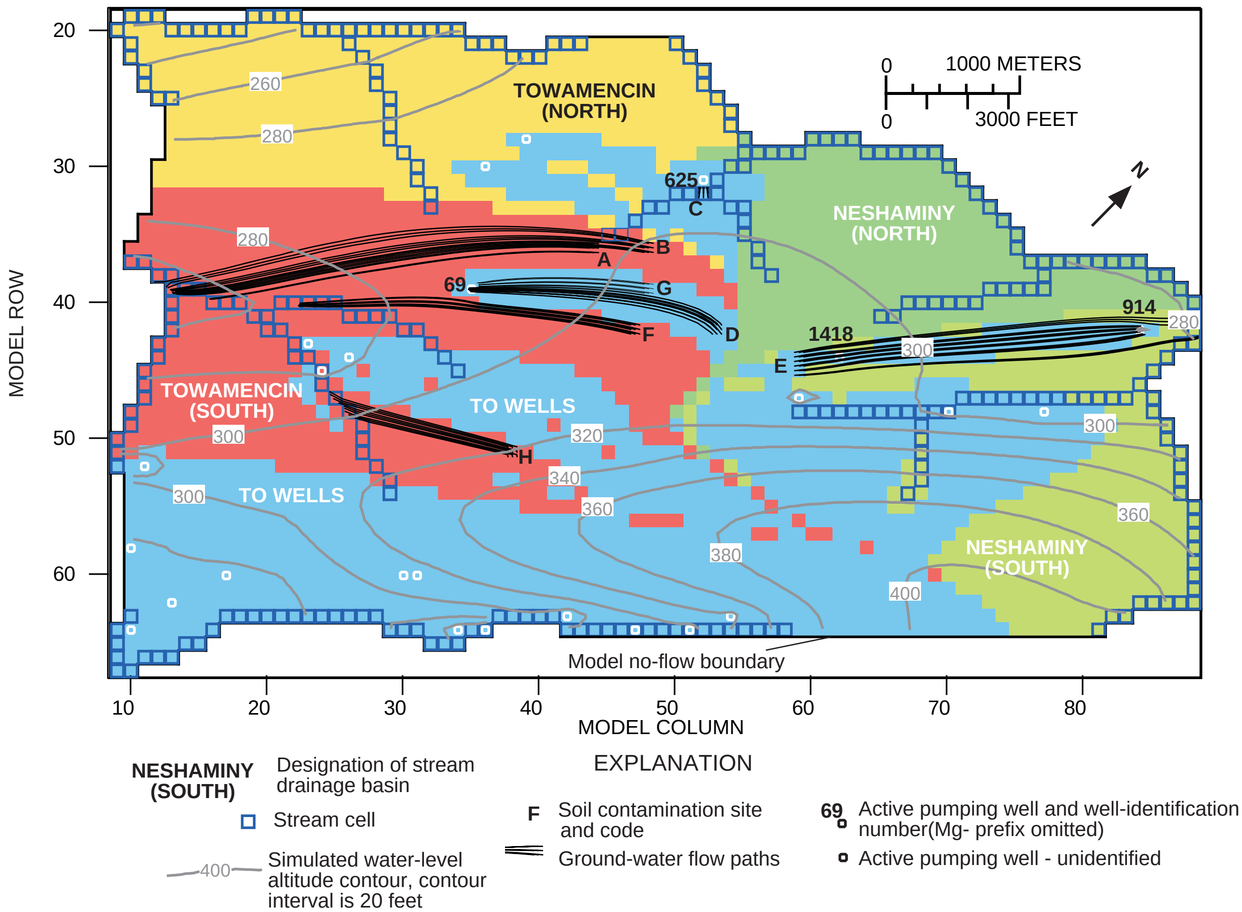


Figure 5.-- Simulated hydraulic head in model layer 2, which represents the upper 328 feet of unweathered, fractured bedrock, and stream and well contributing areas in Lansdale and vicinity for 1997 conditions (modified from Senior and Goode, 1999). Simulated recharge within a colored contributing area discharges to the indicated stream or pumping wells.

SIMULATION OF GROUND-WATER FLOW AT THE LOCAL SCALE

Aquifer tests provide information about aquifer properties and local-scale ground-water flow. Drawdown and changes in ground-water flow on a local scale near a pumping well can be highly variable because of aquifer heterogeneity. Results of aquifer tests in Lansdale indicate that transmissivity differs in both vertical and horizontal directions in the fractured sedimentary rocks that underlie the area (Goode and Senior, 1998; Senior and Goode, 1999). The extent of hydraulic connection between water-bearing fractures is not necessarily related to distance between the fractures but may be related to geologic structure. For example, wells with water-bearing zones located in the projected bed of the pumped interval responded to pumping in aquifer tests (Senior and Goode, 1999). In the Triassic-age sedimentary rocks of the Brunswick Group and the Lockatong Formation, cones of depression caused by pumping have been observed to extend preferentially along strike of bedding planes or in the direction of fracture orientation (Longwill and Wood, 1965).

Local ground-water flow is defined for this report as flow to a single pumping well from a distance of about 1,000 ft (300 m). In this report, data from three multiple-well aquifer tests done in 1997 (QST, Inc. 1998; Senior and Goode, 1999) are used to simulate local-scale ground-water flow for two areas in Lansdale. Detailed local ground-water flow within a borehole between sets of water-bearing fractures is beyond the scope of the simulations in this report.

Approach

A three-dimensional finite-difference numerical model, MODFLOW (Harbaugh and McDonald, 1996), is used to simulate local flow. The model is calibrated using an automatic, nonlinear optimization program, MODFLOWP (Hill, 1992), that minimizes the differences between measured and simulated hydraulic heads and streamflow. MODPATH (Pollock, 1994), a particle-tracking module linked to MODFLOW, is used to calculate and display ground-water-flow pathlines from the output of the flow model. This general approach is the same as that used by Senior and Goode (1999) for a regional-scale model of flow in the Lansdale area.

The model structure is based on a simplified conceptualization of the ground-water-flow system. The weathered and fractured-rock formations were modeled as equivalent porous media, such as unconsolidated granular deposits. Thus, it is assumed that ground-water flow can be described using a three-dimensional flow equation based on Darcy’s Law. In this approach, the hydraulic conductivities used in the model represent the bulk properties of the fractured-rock formations. Water flux, which may pass through only a small fraction of the rock mass occupied by fractures, is simulated as if it were distributed throughout all parts of the formations. Local-scale ground-water flow in fractures and fracture zones is modeled as occurring in stratigraphic beds of high hydraulic conductivity. This approach captures the dominant effect of zones of high hydraulic conductivity on local-scale flow. Detailed characteristics of flow within fractures or within individual beds at scales of a few feet or less are not accurately simulated by the models.

The entire thickness of rock represented by each model layer is assumed to be saturated. This approximation means that the transmissivity (T) of the top model layer is assumed to be independent of the computed hydraulic head. The calibration model MODFLOWP requires this approximation. The model results are relatively insensitive to minor changes in the transmissivity of the top layer because most flow is in the deeper parts of the ground-water system. Where not affected by pumping, the depth to water in the study area commonly is less than 50 ft (15 m) and was less than 30 ft (9 m) in about half of the wells measured in August 1996 (Senior and others, 1998).

The MODFLOWP program calculates optimum values of model parameters, such as recharge rate and hydraulic conductivity, for a particular model structure. The model structure includes all quantitative information that establishes the functional relation between model parameters and predicted heads and streamflow. Although properties of model cells can be specified individually, the approach is to group cells with similar properties into zones with uniform parameters. This approach significantly reduces the

number of model parameters and improves the reliability of parameter estimates. Zones are delineated on the basis of hydrogeologic information.

Limitations and Uncertainties in Predictive Simulations

The contributing areas for pumping wells in the Lansdale area are approximated by the predictive simulations in this report. Although the calibrated models match many of the measured water-level changes during pumping and the regional model reasonably matches overall regional water-level trends, the measurements are not precisely reproduced by the models. Furthermore, steady-state flow under alternative pumping conditions cannot be measured to compare to the model simulations. The actual ground-water flowpaths are likely to be more complex than those shown here because of the highly heterogeneous characteristics of the fractured-rock aquifers, and the flowpaths are likely to change in time because of changing recharge and pumping conditions. The results here can be used to compare the potential effects of alternative ground-water management methods and to indicate general characteristics of contributing areas for these wells. The uncertainties in the predictive simulations could be reduced by more detailed field studies and longer-term aquifer and tracer tests, which are beyond the scope of this report.

North-Central Lansdale Model

A separate local-scale model is constructed for aquifer-test analysis and particle tracking in north-central Lansdale (fig. 4). This area does not contain any streams draining the ground-water system. Hence, the components of the water budget include only recharge, pumping within the area, and fluxes to and from the parts of the aquifer outside the local-scale model area. Furthermore, the aquifer-test results provide information only about a single high-permeability bed (Senior and Goode, 1999). These features allow a small local-scale model to be used that incorporates boundary fluxes determined from the regional-scale model. This approach is computationally efficient but cannot be efficiently used in locations where streams are present within the local-scale model area, or where a regional hydrogeologic structure is simulated beyond the boundaries of the local-scale model.

Aquifer-Test Results

One aquifer test was done at the John Evans and Sons property (Evans) on November 21, 1997 (Senior and Goode, 1999). Well Mg-1609 was pumped for 7.93 hours at rates that ranged from 6 to 10 gal/min (0.38 to 0.63 L/sec) during the early part of the test. The pumping rate was stable at about 9.1 gal/min (0.57 L/sec) from 35 minutes after pumping started until the end of pumping. Water levels were measured in 11 wells (fig. 6) with pressure transducers and electric tapes. Barometric pressure at a nearby site also was recorded with a transducer. The configuration of wells included shallow [about 100 ft (30 m) or less in depth] wells Mg-1533, Mg-1606, Mg-1609 (pumped well), and Mg-1624; an open-hole well (Mg-152) with intermediate [less than about 200 ft (61 m)] and shallow water-bearing zones; intermediate wells Mg-1607, Mg-1666, and Mg-1445; deep [about 300 ft (91 m)] well Mg-1608; and two deep open-hole wells, Mg-618 and Mg-1443, open to a large part of the formation (figs. 7 and 8). Bedding strikes about N45°E and dips about 12° NW in the vicinity of the site (Conger, 1999). Pumping for industrial use occurred intermittently during the aquifer test in well Mg-153 near well Mg-618 (fig. 6).

Positive drawdown during the aquifer test was measured in the pumped well and in 7 of the 10 observation wells (fig. 6). Negative drawdown, probably due in part to barometric effects, was measured in observation wells Mg-618, Mg-1608, and Mg-1624. Drawdown exceeded 0.3 ft (0.1 m) in four observation wells: Mg-1533, a shallow well adjacent to the shallow pumping well (fig. 7 and 8); Mg-152, the next-closest observation well open to shallow and intermediate depths; Mg-1606, a shallow well relatively far from the pumping well but along strike; and Mg-1666, an intermediate depth well downdip of the pumped well but open to the same beds (fig. 7 and 8). Well Mg-1443 is about the same distance from the pumped well as well Mg-152, in the opposite direction along strike, and is open to a large part of the formation. Measured drawdown in well Mg-1443 was less than 0.16 ft (0.05 m), which is less than one-third the drawdown at Mg-152. Drawdown in shallow well Mg-1624 was negative, whereas drawdown in

the adjacent intermediate well Mg-1666 was more than 0.3 ft (0.1 m). These differences in drawdown are consistent with the projection of the pumped beds through the open interval of well Mg-1666 but below that of well Mg-1624 (fig. 7).

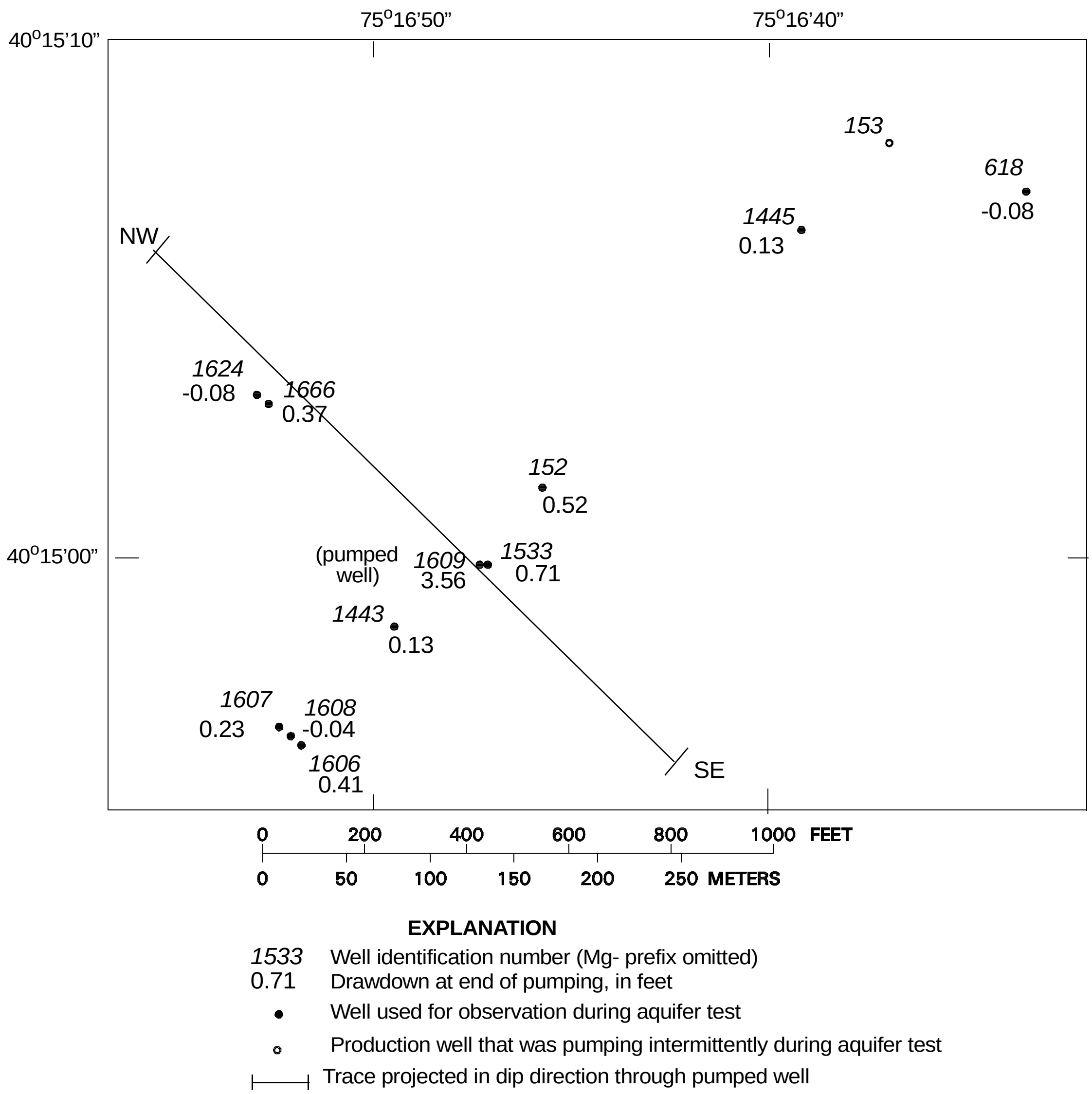


Figure 6.-- Well locations and drawdown at end of pumping well Mg-1609 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (from Senior and Goode, 1999).

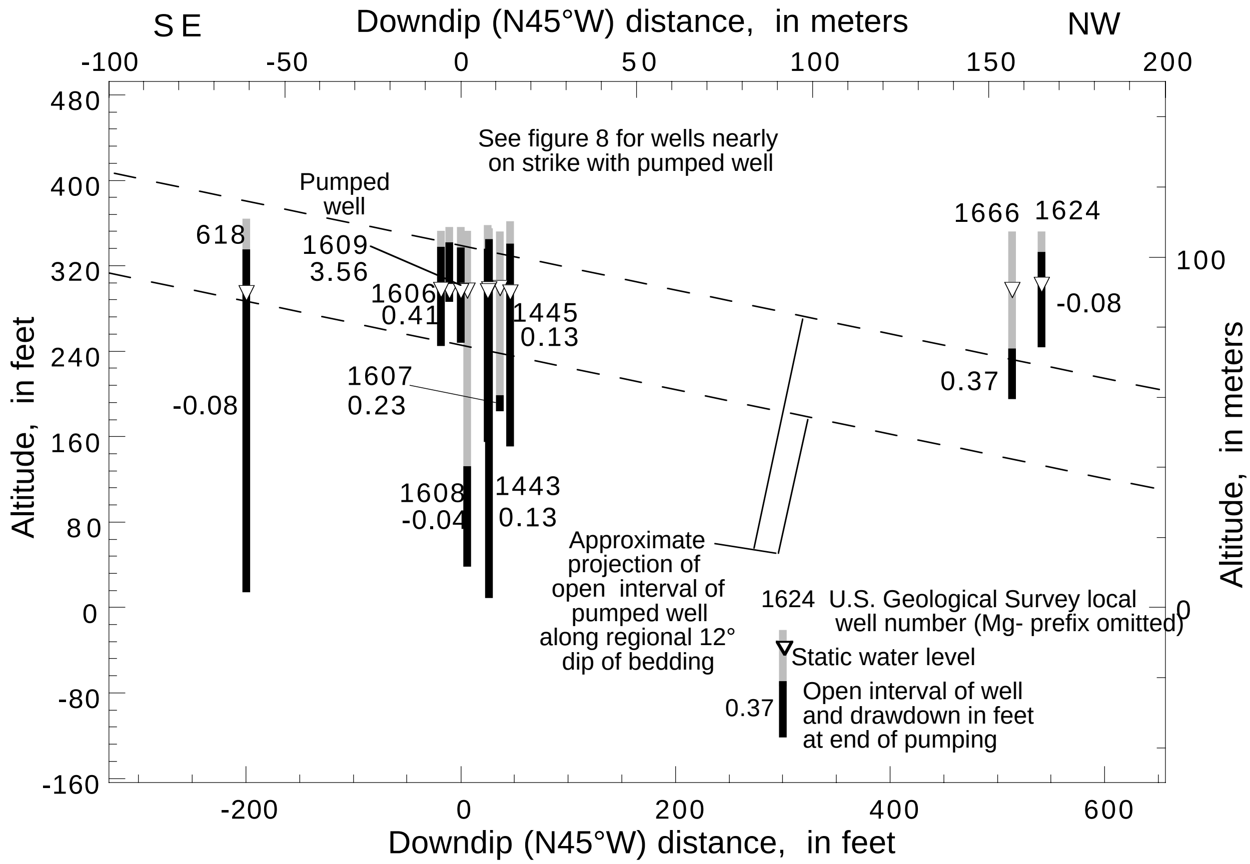


Figure 7.-- Cross-section of open intervals of wells, static depth to water, and drawdown at end of pumping at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction (from Senior and Goode, 1999).

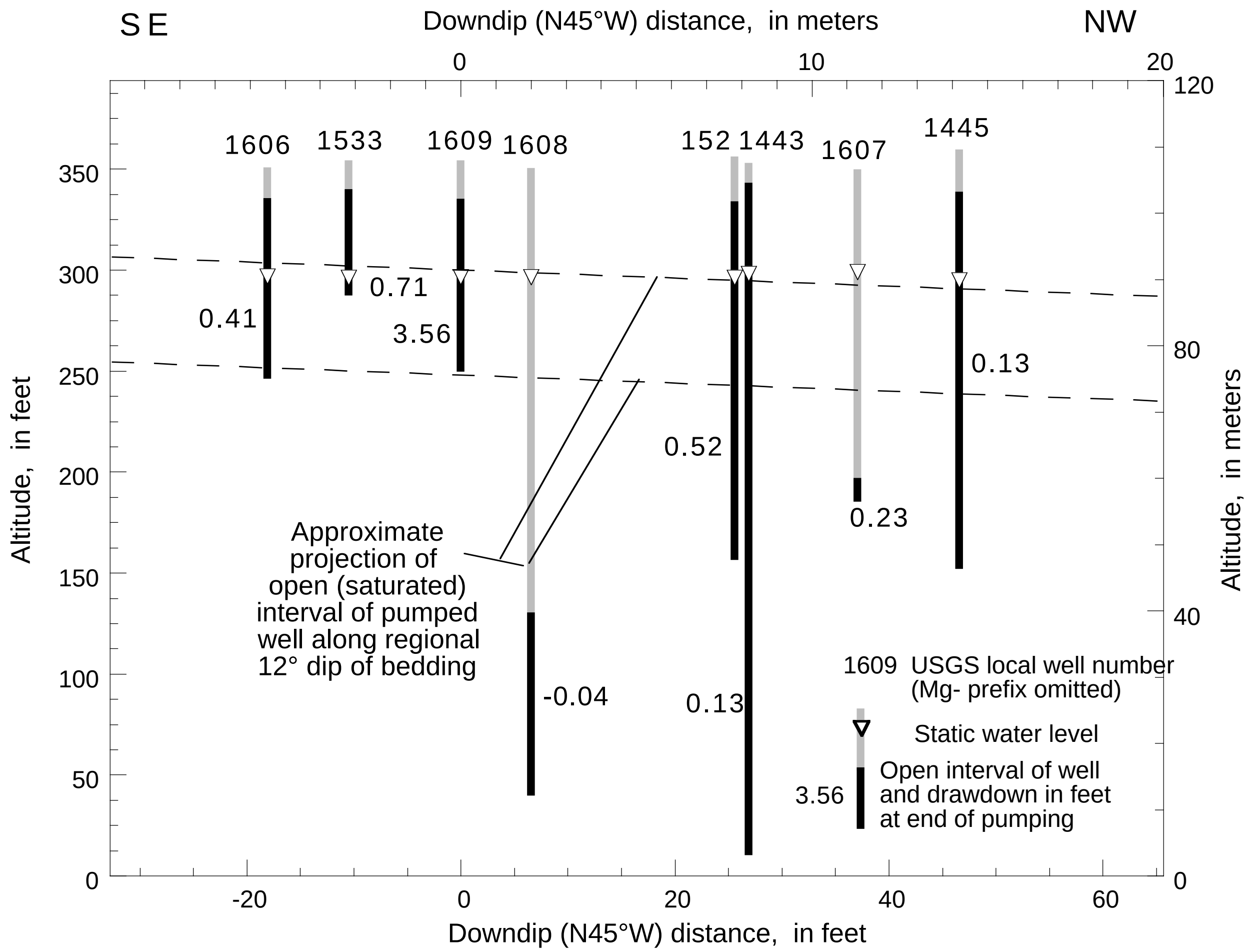


Figure 8.-- Cross-section of open intervals of wells nearly on strike with the pumped well, static depth to water, and drawdown at end of pumping at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction (from Senior and Goode, 1999).

Measured water levels during the aquifer test illustrate the effect of pumping, including variable rates of pumping at the beginning of the test and fluctuations associated with regional water-level trends (fig. 9). The initial pumping rate was up to about 1 gal/min (0.06 L/sec) greater than the long-term average rate, as evidenced by greater drawdown in the pumped well during the first 15 minutes of the test. The water levels in well Mg-1608 (figs. 8 and 9) are representative of the other two observation wells (Mg-618 and Mg-1624) that did not respond to pumping. The water level in well Mg-1608 did respond to changes in barometric pressure (fig. 9) and rose about 0.04 ft (0.01 m) over the pumping period of the test. Water levels in well Mg-1445 apparently responded to pumping in well Mg-1609 but also responded strongly to other pumping in the area. Other pumping also resulted in minor water-level changes in the other observation wells. For wells included in the aquifer-test analysis, drawdown was not corrected for the apparently small effects of barometric pressure decrease or other pumping wells. The recovery of water levels in the pumped well is similar to that reported for many pumping tests in the Lansdale area (Goode and Senior, 1998). A very rapid recovery of more than 75 percent of the drawdown at the end of pumping was followed by a much more gradual recovery to the static water level.

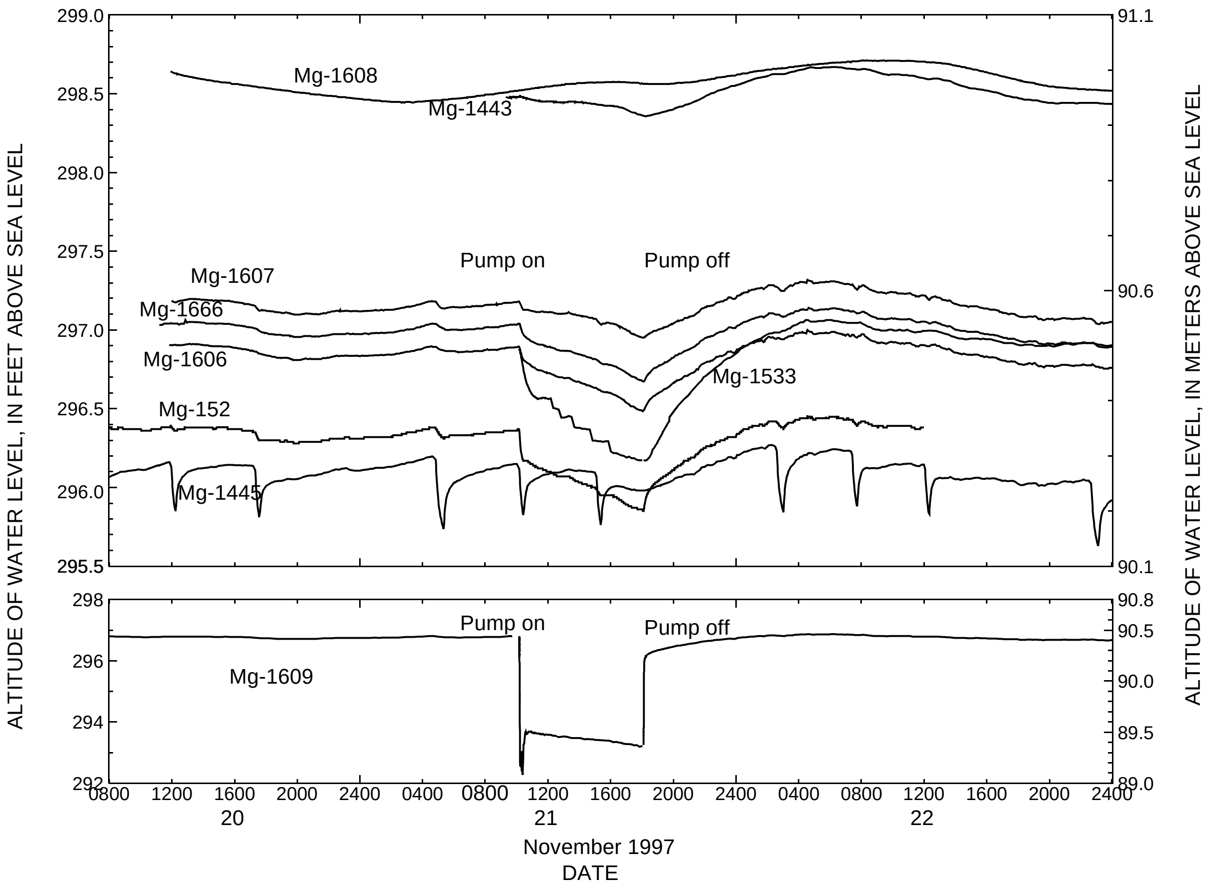


Figure 9.-- Measured water levels at the John Evans and Sons property in north-central Lansdale, Pa., November 20-22, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours on November 21 (from Senior and Goode, 1999).

Senior and Goode (1999) matched drawdown in four observation wells using the two-aquifer analytical model of Neuman and Witherspoon (1969) to estimate transmissivity (T), storage coefficient (S), hydraulic conductivity (K), and specific storage (S_s) (fig. 10). These four wells had the largest measured drawdowns. The two-aquifer model matches the measured drawdown in these four wells better than either the isotropic Theis model or the anisotropic single-aquifer model (Papadopulos, 1965). Smaller drawdown at several other observation wells could not be matched by using this conceptual model. The estimated hydraulic properties from this match are T_1 = 1,300 ft^2/d (122 m^2/d), $S_1 = 8 \times 10^{-5}$ for the pumped 'aquifer' or network of fractures; T_2 = 15 ft^2/d (1.4 m^2/d), $S_2 = 8 \times 10^{-5}$ for the unpumped 'aquifer'; and K_v = 0.044 ft/d (0.013 m/d), and $S_s = 1 \times 10^{-6}$ /ft (3×10^{-6} /m) for the low-permeability unit separating the two aquifers. These results are consistent with the results of aquifer interval-isolation tests (Senior and Goode, 1999) in that the vertical hydraulic conductivity is very low for bedrock between high-permeability zones oriented along bedding.

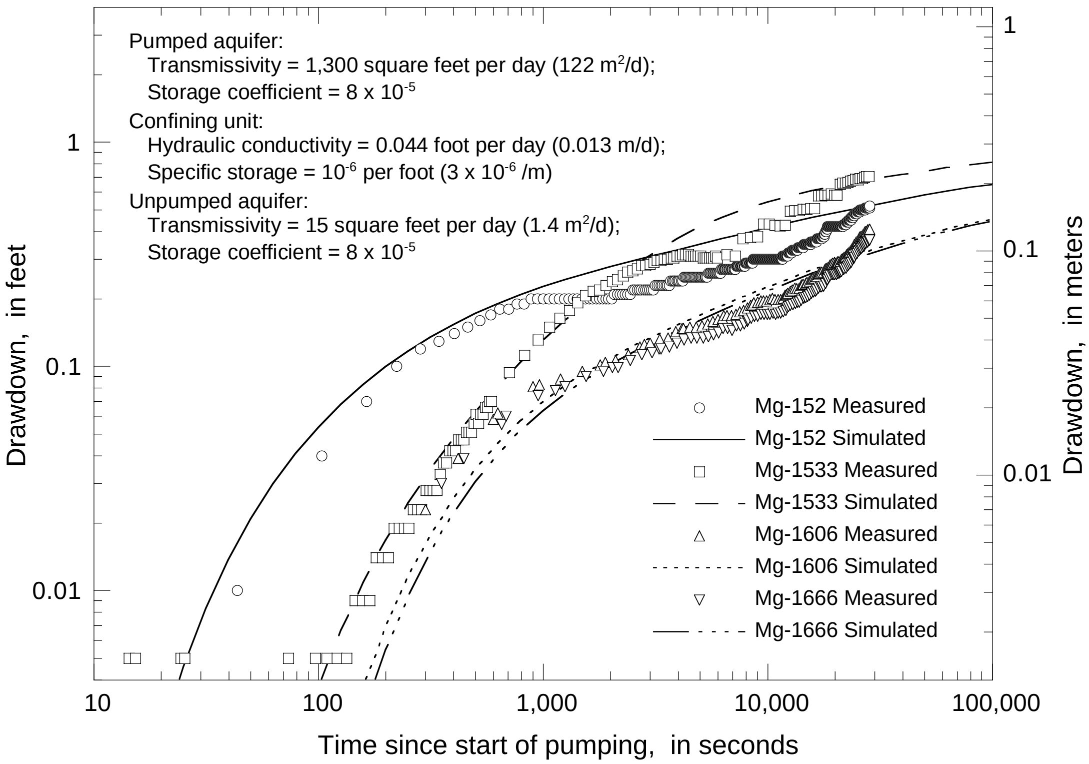


Figure 10.-- Measured and simulated drawdown, using two-aquifer model of Neuman and Witherspoon (1969), in wells Mg-67, Mg-80, Mg-163, and Mg-1666 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (from Senior and Goode, 1999).

Model Structure and Boundary Conditions

The local-scale model of ground-water flow in the north-central part of Lansdale (fig. 4) is based on the regional-scale model. The thickness of the entire aquifer is divided into three major layers and includes a dipping high-permeability bed within the bedrock (fig. 11). The hydrogeologic layers are divided into 12 layers for model computations. The soil zone is represented by model layer 1 and is uniformly 16.4 ft (5 m) thick. The upper weathered part of the bedrock is represented by model layer 2 and is uniformly 16.4 ft (5 m) thick. The unweathered fractured bedrock is represented by model layers 3 through 12; the thickness of the layers increases progressively with depth from 16.4 to 82 ft (5 to 25 m). The bedrock layers in this model correspond to layers 2 and 3 of the regional-scale model of Senior and Goode (1999). The pumped well is open to a dipping, high-permeability bed that extends throughout the area of the local model. The vertical position of this bed depends on the local dip and strike. Hence, this pumped bed is represented by a stair-step configuration of high-permeability cells occurring in all model layers 2-12 (fig. 11).

Areally, model rows are aligned with the strike of the local stratigraphy (fig. 4). The horizontal dimensions of model grid cells range from 9.8 by 9.8 ft (3 by 3 m) at the pumping well to 65.6 by 65.6 ft (20 by 20 m) for cells more than about 164 ft (50 m) from the pumping well.

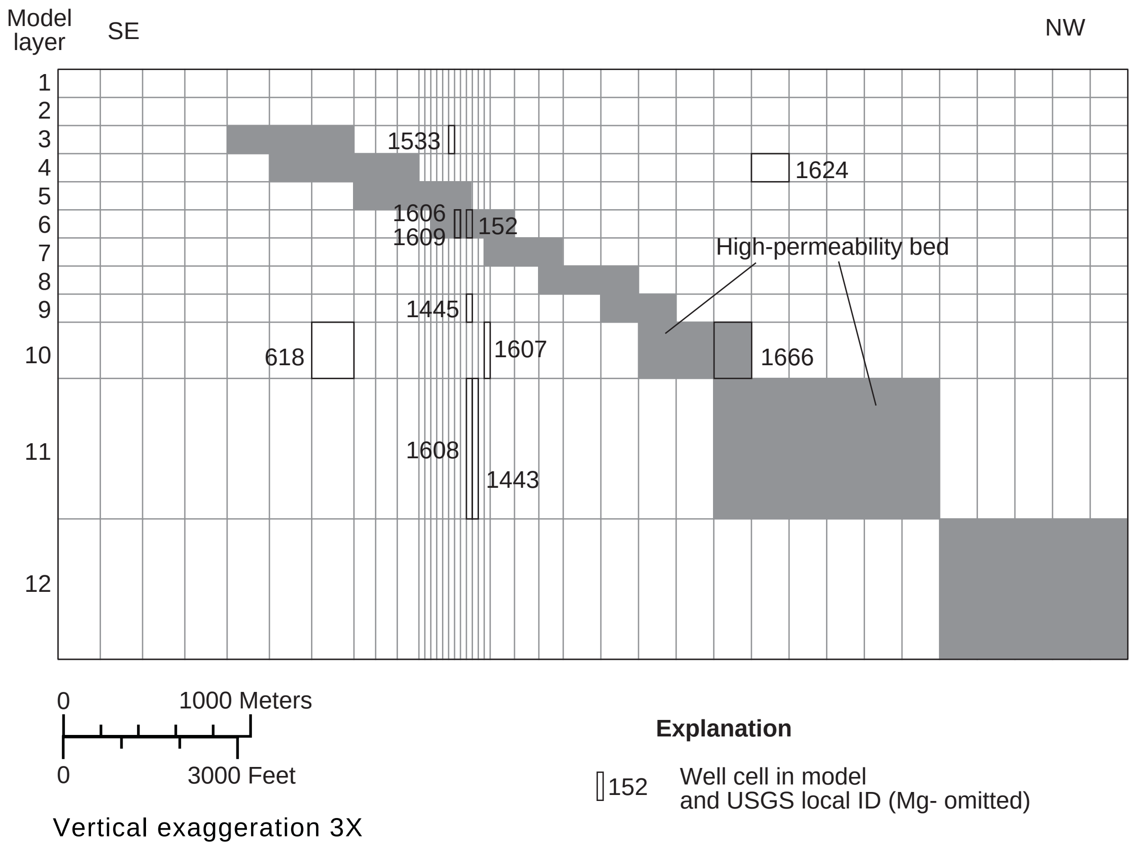


Figure 11.-- Cross section of local-scale model of ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa. Well locations in the model are indicated; wells 1606 and 1609 are collocated in the cross section but are in different model columns. See fig. 4 for location of local-scale model.

Two different model configurations (fig. 12) are used: (a) a local-scale model for calibration to aquifer-test results; and (b) a local-scale model with overlap areas beyond the boundary of the local-scale model for ground-water-flowpath simulation. The local-scale model is calibrated to the aquifer-test results assuming that there is insignificant interaction between the local model and regional flow. For model calibration to the aquifer-test results, initial heads are specified as zero throughout the local model domain, and no-flow conditions are applied along all model boundaries.

Ground-water flowpaths under steady-state conditions are simulated with overlap areas around the local model to accommodate the specified flux boundary conditions from the regional-scale model. The outer five columns on the SW and NE ends of the model and the outer three rows on the NW and SE edges of the model are overlap areas beyond the boundaries of the local-model domain (fig. 12). In the regional model, the bedrock is homogeneous; hence, flow is nearly evenly distributed vertically. In the local-scale model, however, some of the cells along the boundary between the local and regional models have high permeability, and others have low permeability. Rather than setting a similar specified flux in cells with different properties, the specified flux is applied on the outside of the model overlap region (fig. 12). In this overlap region, properties are uniform. The specified boundary flux from the regional-scale model is applied on the outside of the overlap area. The same overall flux occurs at the local / regional boundary, but most of the flow occurs where high-permeability cells are located. The specified flux values are determined by regional-scale model simulation using pumping rates of the “1997” simulation of Senior and Goode (1999) with the additional specified pumping at well Mg-1609.

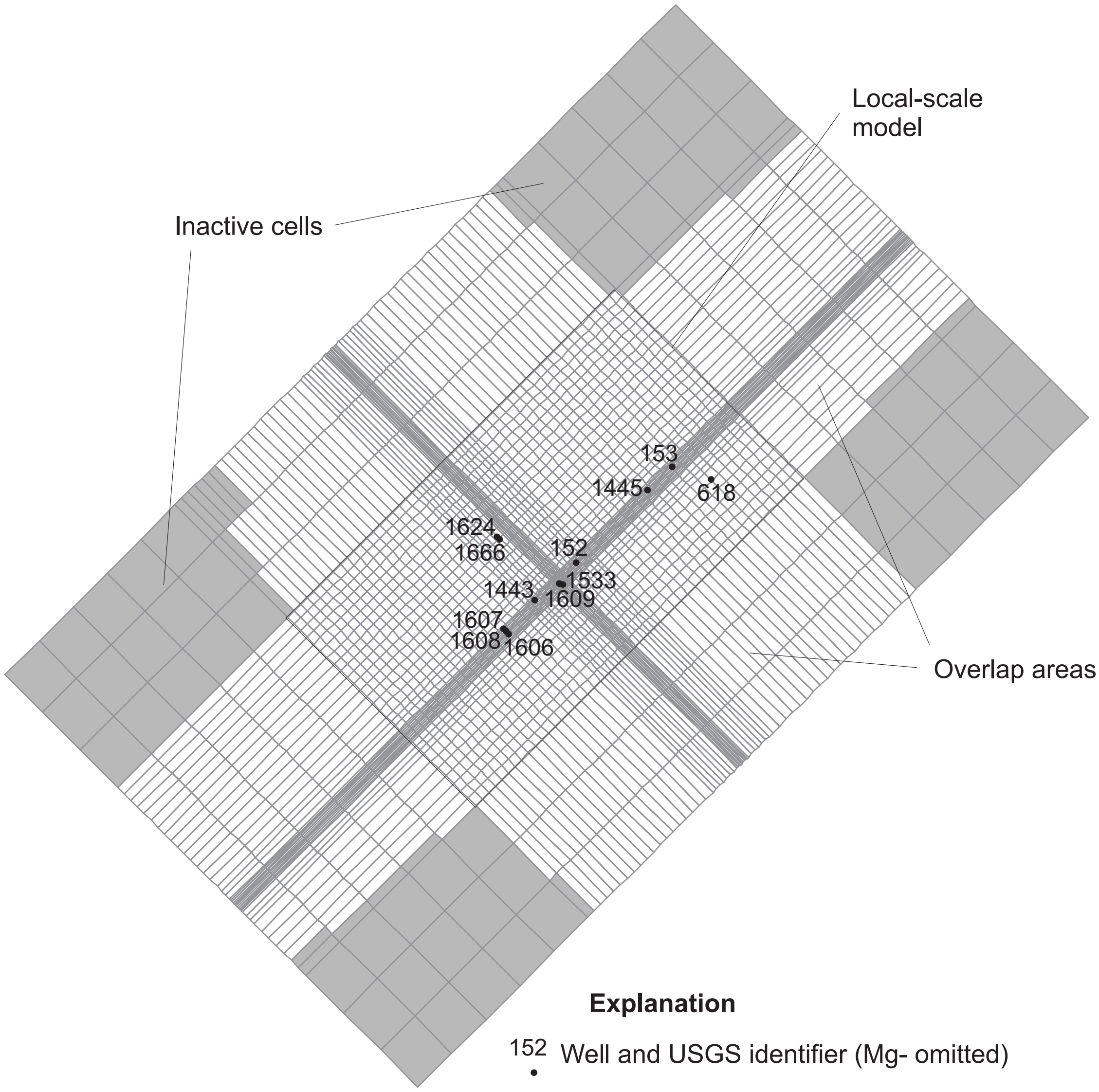


Figure 12.-- Well location and horizontal grid configuration for simulation of local-scale ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa. showing local-scale grid and overlap areas for incorporation of regional-model flux boundary conditions.

Aquifer-Test Simulation

The local-scale model is calibrated by simulation of drawdown during the aquifer test of November 1997. The model is calibrated to drawdown measured in the same four wells used for the analytical model analysis summarized in the Aquifer-Test Results section. Three of the wells used in the analysis are in high-permeability model cells, representing the pumped bed, and the fourth well is in a cell that is within the low-permeability part of the aquifer, not in the pumped bed.

MODFLOWP automatically determines the optimum values of model parameters (hydraulic conductivity and storage coefficient) that yield the minimum sum of squared errors (table 3). Model errors are the difference between simulated and measured drawdown. This procedure is similar to that of matching analytical models to the measured drawdown (e.g. Senior and Goode, 1999, p. 55), except that a numerical model of flow is used here. The results of simulations obtained here with a three-dimensional numerical model are similar to the results obtained by Senior and Goode (1999, p. 67) using a two-aquifer analytical model. The transmissivity of the pumped bed estimated here, 1,900 ft^2/d, is about 45 percent higher than the transmissivity of the pumped aquifer reported by Senior and Goode (1999), 1,300 ft^2/d. The hydraulic conductivity of the rest of the bedrock, 0.05 ft/d, is only slightly larger than the hydraulic conductivity of the low-permeability unit separating the aquifers in the analytical model (Senior and Goode, 1999), 0.044 ft/d. The storage parameters agree within a factor of 2. The analytical model included a second unpumped aquifer that does not have a corresponding part in the numerical model used here.

The calibrated model can approximately simulate measured drawdowns measured during the aquifer test (fig. 13). Compared to the analytical match using the two-aquifer model by Senior and Goode (1999, fig. 10), this model does not match early drawdown as well but more closely matches the rate of drawdown increase at late time (fig. 13). The best fit of the model is obtained with high K in the pumped bed and low K in the lower permeability rock (table 3). The weathered rock and soil also have low K.

Table 3.-- Optimum and approximate, individual, 95-percent confidence-interval values for hydraulic conductivity and specific storage for calibrated simulation of ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa.

[ft^2/d, feet squared per day; ft/d, foot per day]

Parameter	Units	Optimum value	Approximate, individual, 95-percent confidence interval: Lower value	Approximate, individual, 95-percent confidence interval: Upper value
Pumped bed transmissivity	ft^2/d	1,900	500	2,300
Bulk rock hydraulic conductivity	ft/d	0.05	.033	.077
Pumped bed storage coefficient	-	4.0×10^{-5}	2.3×10^{-5}	7.0×10^{-5}
Bulk rock specific storage	per foot	1.5×10^{-6}	9.9×10^{-7}	2.3×10^{-6}

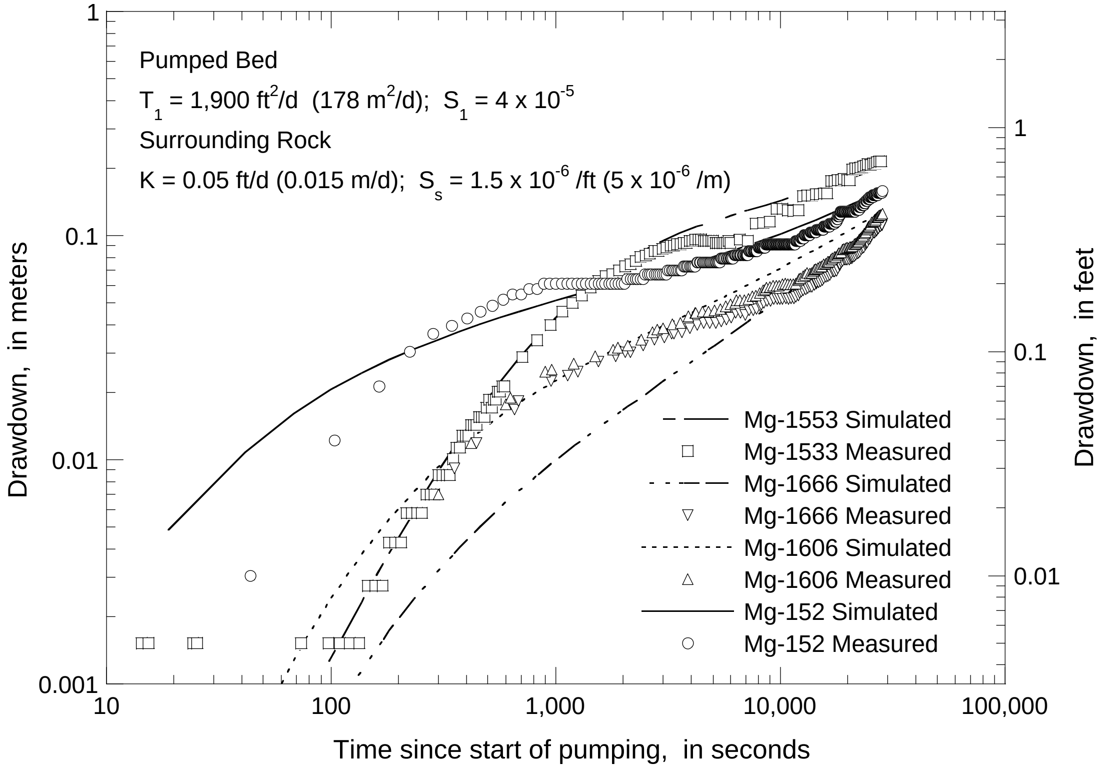


Figure 13.-- Measured and simulated drawdown, using local-scale flow model, in wells Mg-67, Mg-80, Mg-163, and Mg-1666 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (measured drawdown from Senior and Goode, 1999).

The spatial pattern of drawdown within a horizontal model layer is characteristic of anisotropy (fig. 14), although the model is configured to represent a heterogeneous isotropic aquifer. Within a model layer, only some of the cells are located in the high-hydraulic-conductivity pumped bed. Head gradients within these cells are small because the hydraulic conductivity is high. The configuration of cells with high hydraulic conductivity along rows leads to the elongated drawdown contours, similar to results in anisotropic homogeneous aquifers.

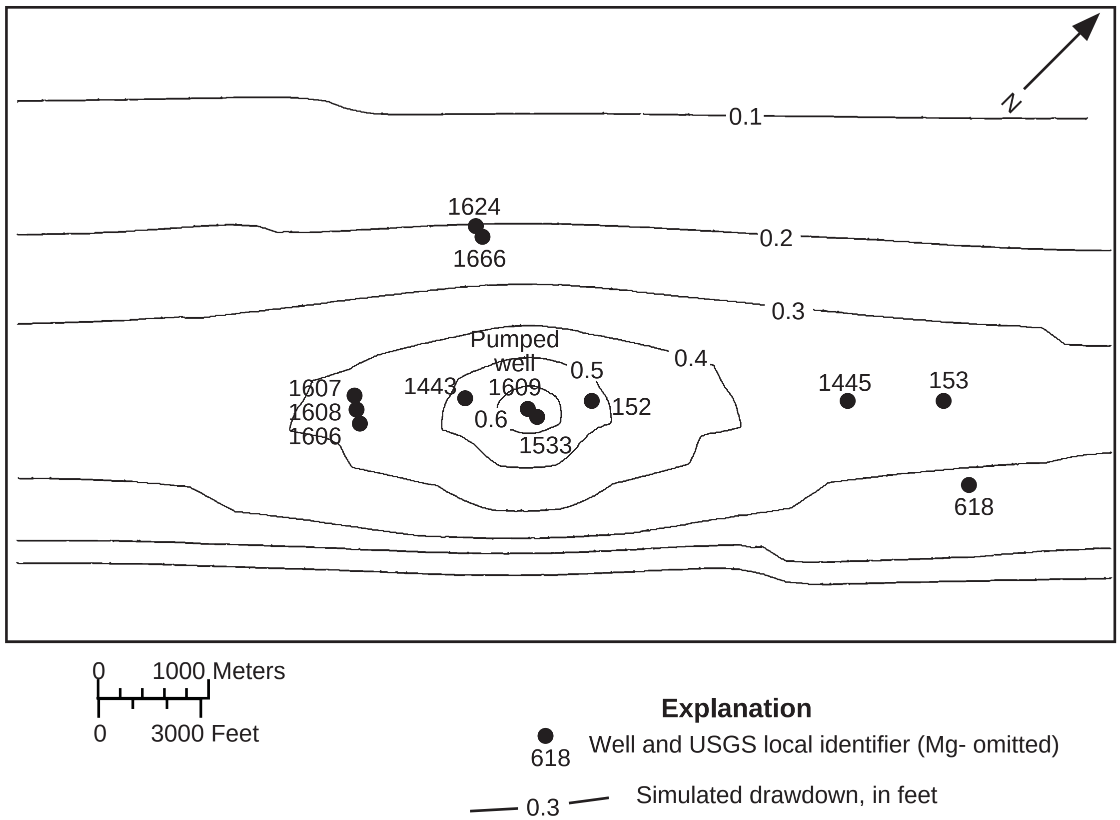


Figure 14.-- Well locations and simulated drawdown in model layer 6 (containing the pumped well) after 7.93 hours of pumping well Mg-1609 at a rate of 9.1 gallons per minute at the John Evans and Sons property in north-central Lansdale, Pa.

Effect of Pumping on Ground-Water Flowpaths

A steady-state-flow field is simulated in the local-scale model using boundary fluxes from the regional-scale model with the specified pumping in well Mg-1609. Regionally, flow into the local-scale model occurs along the SE boundary; most flow out crosses the SW and NE boundaries. When pumping at 10 gal / min, about 12 percent of the inflow to the local model discharges to the pumping well.

The source areas for water pumped from well Mg-1609 are illustrated by results of a particle tracking simulation. Particle paths are backtracked from the pumping well to the local-scale model boundaries using MODPATH (Pollock, 1994). About 56 percent of the particles originate at the water table in the local-model domain (fig. 15). The remaining particles cross the inflow boundary along the SE edge of the model and originate outside the local-model domain. Deeper particles on this edge of the model probably originate at the water table far from the pumped well. Any recharge more than about 200 ft from the pumping well in the downgradient (NW) direction is not captured by the pumping well.

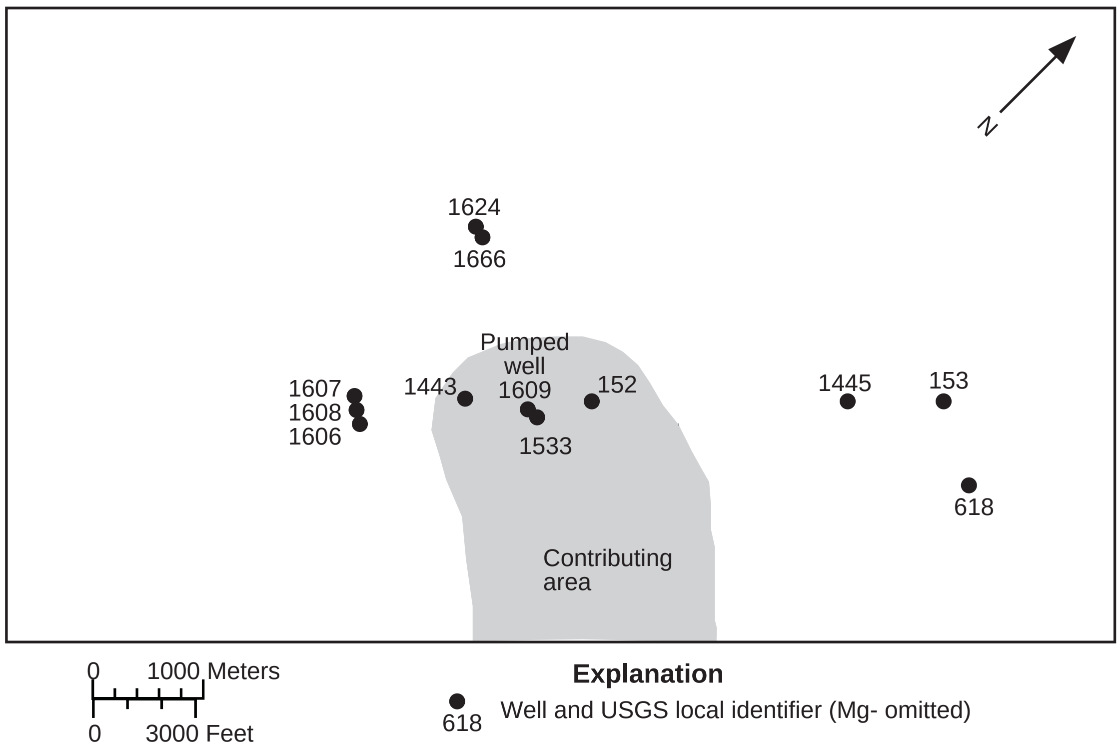


Figure 15.-- Simulated contributing area for well Mg-1609 pumping at a rate of 10 gallons per minute at the John Evans and Sons property in north-central Lansdale, Pa. See figure 4 for location of the local-scale model.

Northwestern Lansdale Models

Local-scale models of ground-water flow in northwestern Lansdale are embedded within the regional-scale flow model of Senior and Goode (1999). This approach is used because the two local-scale models are linked by high-permeability beds that extend regionally and because streams traverse the local-scale model areas. This approach is computationally less efficient than separate local and regional models but allows ground-water fluxes between the local-scale model areas and the regional model to be automatically simulated. Furthermore, it allows a simultaneous calibration of the model using results from the two separate aquifer tests.

Aquifer tests were conducted at the J.W. Rex Co. property (Rex) in October 1997 and at the Keystone Hydraulics property (Keystone) in November 1997. Although these locations are far enough apart that the aquifer test at one did not appreciably affect water levels at the other during the tests, they are within a mile of each other and are nearly along strike. Calibrating a single model to both aquifer tests yields more representative hydraulic properties and facilitates simulation of flowpaths at the local as well as larger scale.

Aquifer-Test Results

Keystone Hydraulics site

One aquifer test was done at Keystone on November 18, 1987 (Senior and Goode, 1999). Well Mg-1610 was pumped for 8.05 hours at rates that ranged from 8.1 to 15 gal / min (0.51 to 0.95 L / sec) during the early part of the test. The pumping rate was stable at about 10 gal / min (0.63 L / sec) from 42 minutes after pumping started until the end of pumping. Water levels were measured in eight wells (fig. 16) with pressure transducers and electric tape. The configuration of wells included shallow [less than 100 ft (30m)] wells Mg-1611 and Mg-1620; intermediate-depth [up to 190 ft (60 m)] wells Mg-1610 (pumped well) and Mg-1619; and several deep [more than 270 ft (82 m)] open-hole wells (Mg-67, Mg-80, Mg-163, and Mg-164) (fig. 17). The observation wells were updip and along strike from the pumped well. Bedding at Keystone strikes about N57°E and dips about 8° to the northwest (Conger, 1999).

Positive drawdown during the aquifer test was measured in all wells but Mg-164 (fig. 16). Drawdown exceeded 0.3 ft (0.09 m) in three observation wells that were among the closest to and updip of the pumped well including Mg-1611, a shallow well within 25 ft (7.6 m) of the intermediate depth pumped well; Mg-80, an open-hole deep well with 138 ft (42 m) of casing and within 153 ft (46.6 m) of the pumped well; and Mg-1620, a shallow well within 365 ft (111m) of the pumped well. Well Mg-1611 is not open to the projected pumped interval. Large drawdown at this well may be caused by local high permeability outside the projected bed. Although the primary water-bearing zone in well Mg-80 is about 30 ft (9.1 m) below the projected dip of bedding through the pumped zone, aquifer interval-isolation testing indicated that this water-bearing zone in well Mg-80 may be hydraulically connected to shallower zones outside the borehole. Shallow well Mg-1620 intersects the projected dip of bedding through the pumped zone (fig. 17). Well Mg-1619 is at a similar distance from the pumped well as well Mg-1620 and is within 25 ft (7.6 m) of well Mg-1620, yet drawdown in well Mg-1619 is only 0.13 ft (0.04 m). Well Mg-1619 is open to beds that are projected to be below the pumped bed (fig. 17). Water levels in well Mg-163, approximately along strike with the pumped well, were drawn down by more than 0.14 ft (0.05 m), whereas water levels in well Mg-164, at a similar radial distance but further updip, were not affected by pumping.

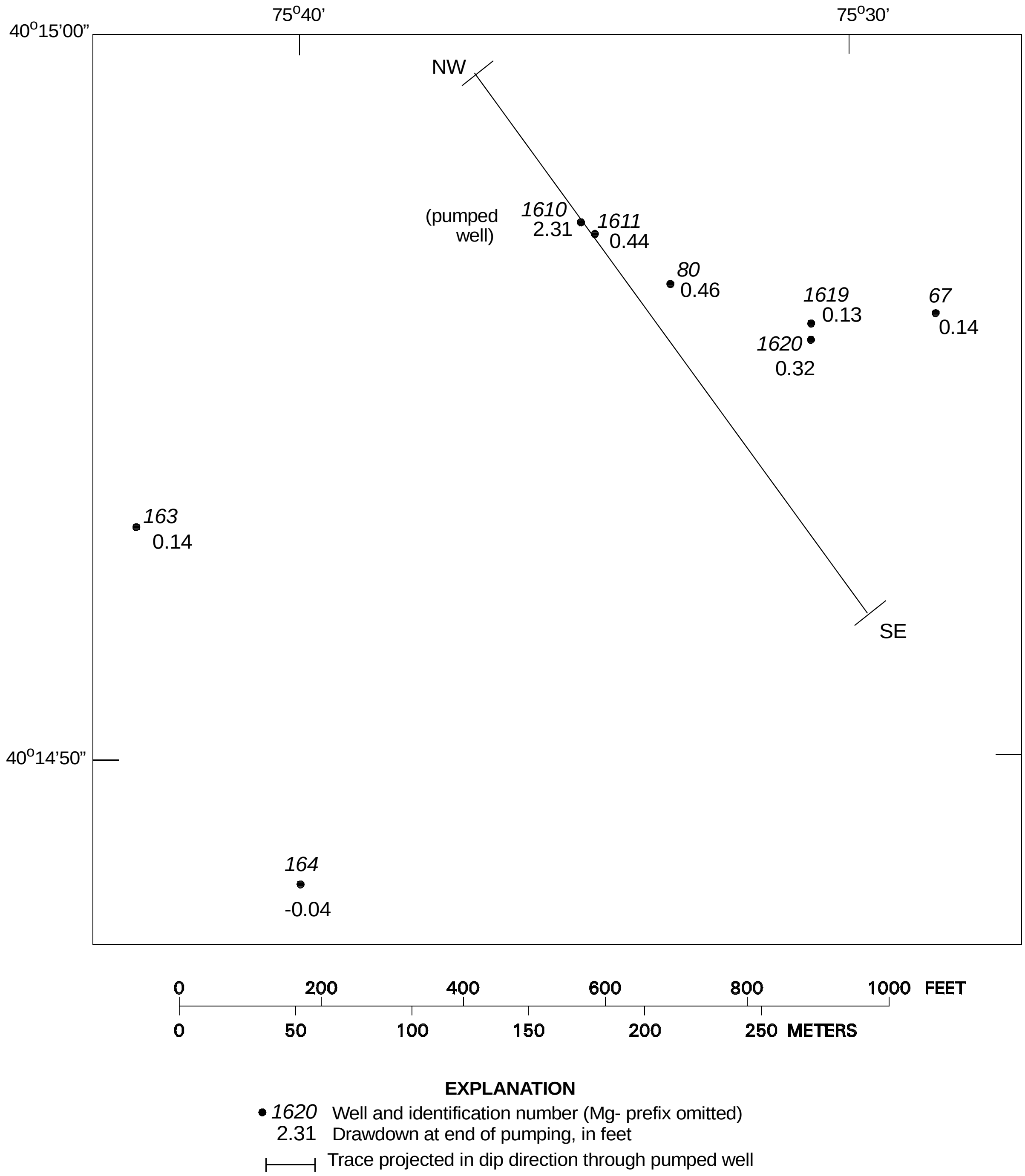


Figure 16.-- Well locations and drawdown at end of pumping well Mg-1610 at the Keystone Hydraulics property in northwestern Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours (from Senior and Goode, 1999).

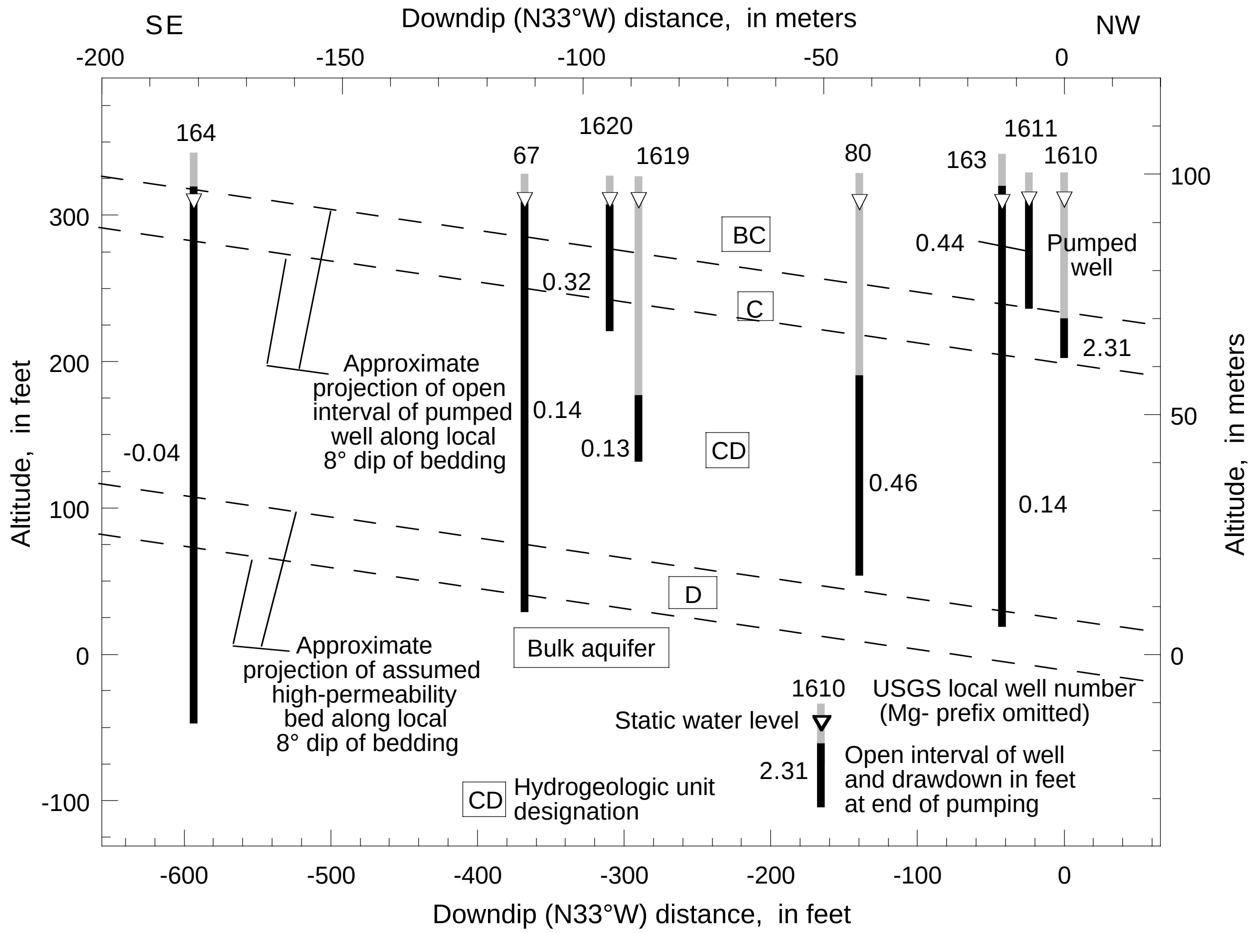


Figure 17.-- Cross-section of open intervals of wells, static depth to water, and drawdown at end of pumping at the Keystone Hydraulics property in northwestern Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours. Also shown is the conceptual model of dipping high-permeability (C, D) and low-permeability (BC, CD) beds. High and low-permeability beds stratigraphically below D are combined and designated “Bulk aquifer.” All wells are projected onto a vertical plane parallel to the dip direction (modified from Senior and Goode, 1999).

Measured water levels during the aquifer test illustrate the effect of pumping, including variable rates of pumping at the beginning of the test and fluctuations associated with regional water-level trends (fig. 18). Decreases in barometric pressure resulted in corresponding increases in water levels in wells during the aquifer test. Because the drawdowns resulting from pumping were small, Senior and Goode (1999) used analytical aquifer-test models to remove the effect of the barometric pressure changes prior to analysis of drawdown.

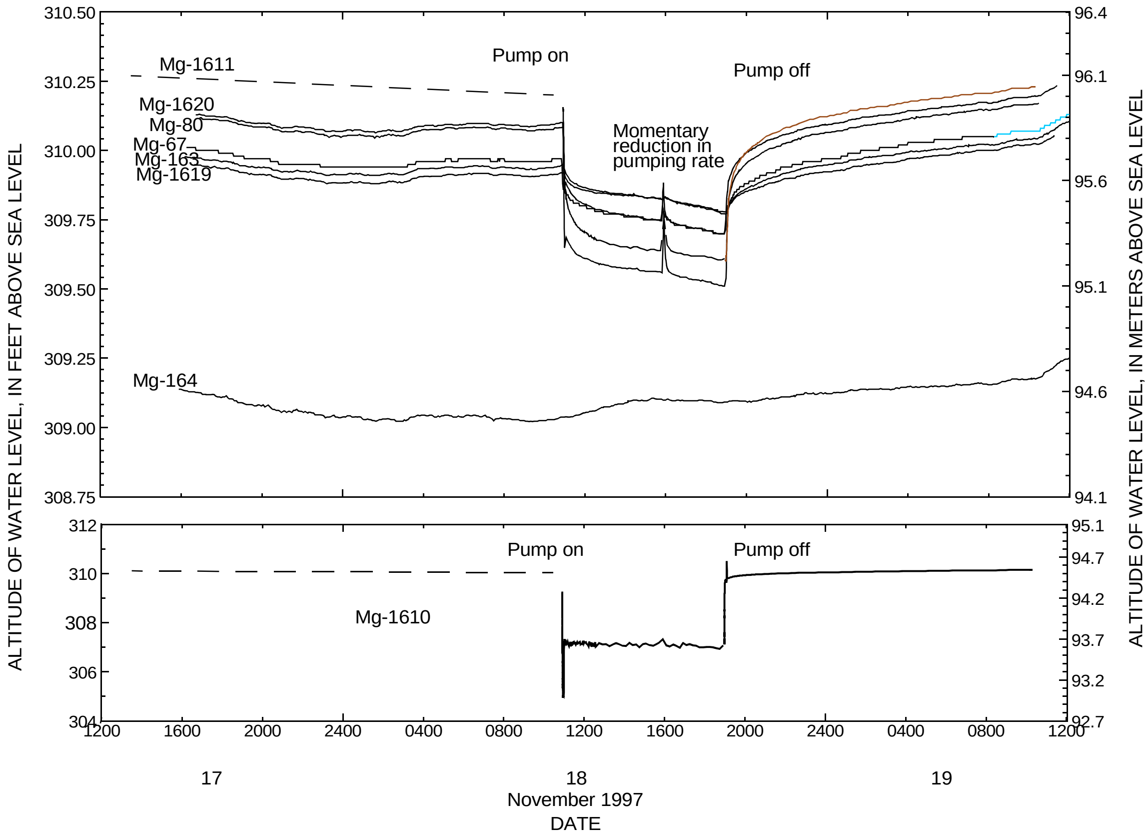


Figure 18.-- Measured water levels at the Keystone Hydraulics property in northwestern Lansdale, Pa., November 17-19, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours on November 18 (from Senior and Goode, 1999).

Senior and Goode (1999) selected drawdown in four observation wells for analysis using the anisotropic single-aquifer model of Papadopulos (1965). Of the six observation wells with positive drawdown, drawdown in two wells was not matched. Well Mg-1611 is very close to the pumping well but was drawn down less than more distant wells, and the well is not open to the projected pumped bed (fig. 17). Well Mg-1619 was drawn down less than half as much as the nearby well Mg-1620, and it also is not open to the projected pumped bed. Drawdown in these wells cannot be matched by a single-aquifer model because in such a model all observation wells are assumed to be completed within the pumped aquifer. Senior and Goode (1999) did not include these wells in their analysis in order to use the directional variability of drawdown in the pumped bed to estimate large-scale anisotropy. Well Mg-80 was included in the analysis even though it also is open to bedrock outside the projected pumped interval. The measured drawdown and aquifer-isolation test results suggest that this well is hydraulically connected to the pumped interval (Senior and Goode, 1999).

Senior and Goode (1999) matched drawdown in four observation wells using the single-aquifer anisotropic model of Papadopulos (1965) (fig. 19). The response of anisotropic aquifers to aquifer tests include larger drawdowns in one direction than in another for similar distances from the pumped well. The early-time part of the measured drawdown was not matched because the pumping rate was variable for about the first 42 minutes of pumping. The estimated hydraulic properties from this match were: T_{max} = 10,700 ft^2/d (990 m^2/d); T_{min} = 520 ft^2/d (48 m^2/d); horizontal angle of maximum T, θ_{max} = N51°W; and S = 3 x 10^{-5}. The non-directional geometric-mean transmissivity was 2,300 ft^2/d (220 m^2/d). These aquifer-test results from this match represent a preferred flow direction within the pumped bed that is oriented in the dip direction (about N33°W). Previous aquifer-test results in the similar Passaic Formation (Carleton and others, 1999) present a preferred flow direction oriented in the strike direction.

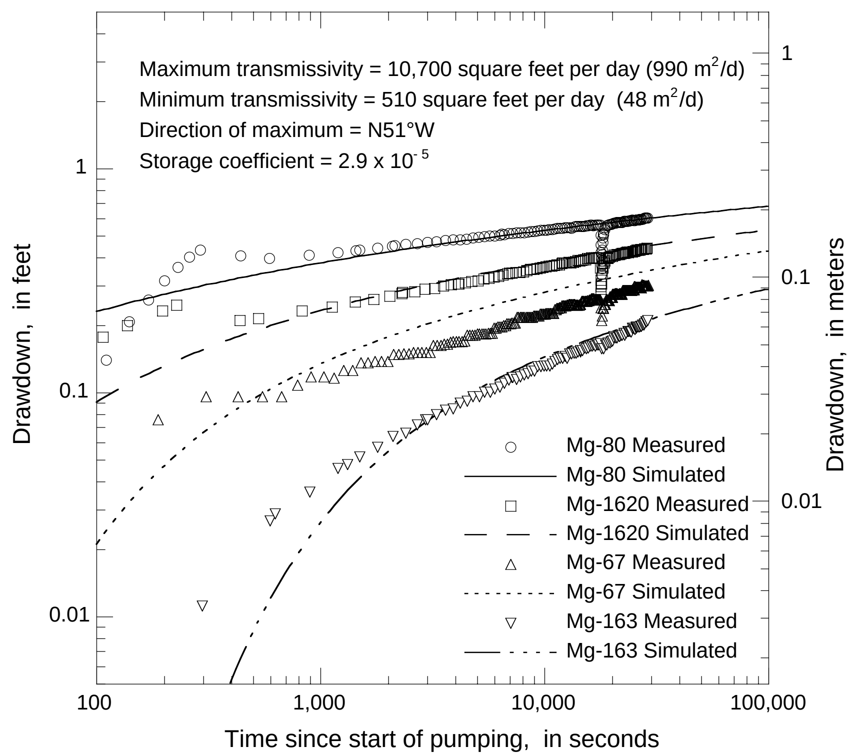


Figure 19.-- Measured and simulated drawdown, using anisotropic model of Papadopulos (1965), in wells Mg-67, Mg-80, Mg-163, and Mg-1620 at the Keystone Hydraulics property in north-western Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours (from Senior and Goode, 1999).

Rex site

Aquifer and step tests at Rex were done by QST Environmental, Inc. (1998). Production well Mg-625 was pumped at a rate of about 30 gal / min for 60 hours from October 24-27, 1997, and then shut down for 13 hours. A step test was conducted at the beginning of the 60-hour pumping period. This well is normally pumped continuously for industrial water supply. Water levels in the pumped well and 10 other wells, including Mg-82, Mg-157, Mg-1441, Mg-624, Mg-1639, Mg-1640, Mg-1641, Mg-1615, Mg-1617, and Mg-1665 (fig. 20), were measured during the tests. Water-level changes were observed in all wells. Recovery at the end of the shutdown period was greatest [12 ft (3.7 m)] in observation well Mg-1639. Well Mg-1639 is the closest to the pumped well. Well Mg-1640 is within 10 ft (3 m) of well Mg-1639 but is shallower than well Mg-1639 and had much less recovery [3.0 ft (0.9 m)]. Aquifer interval-isolation tests indicate little hydraulic connection between well Mg-1639 and the screened interval of well Mg-1640 (QST Inc., 1998; Senior and Goode, 1999). The downward vertical flow observed during geophysical logging prior to the aquifer tests indicates that well Mg-1639 is directly influenced by pumping in production well Mg-625. Estimates of hydraulic properties were determined from analysis of drawdown data, assuming an isotropic aquifer. Transmissivity ranged from 160 to 665 ft^2/d (14.5 - 61.8 m^2/d) and storage ranged from about $2x10^{-5}$ to $4x10^{-3}$ (QST Environmental Inc., 1998). The transmissivity values from this test are similar to a value of 330 ft^2/d (31 m^2/d) estimated from an earlier test (Goode and Senior, 1998).

The extent of water-level changes in observation wells are controlled, in part, by the hydrostratigraphic locations of open intervals (fig. 21). Wells that are open to the same hydrostratigraphic horizon as the pumped interval have the largest water-level changes.

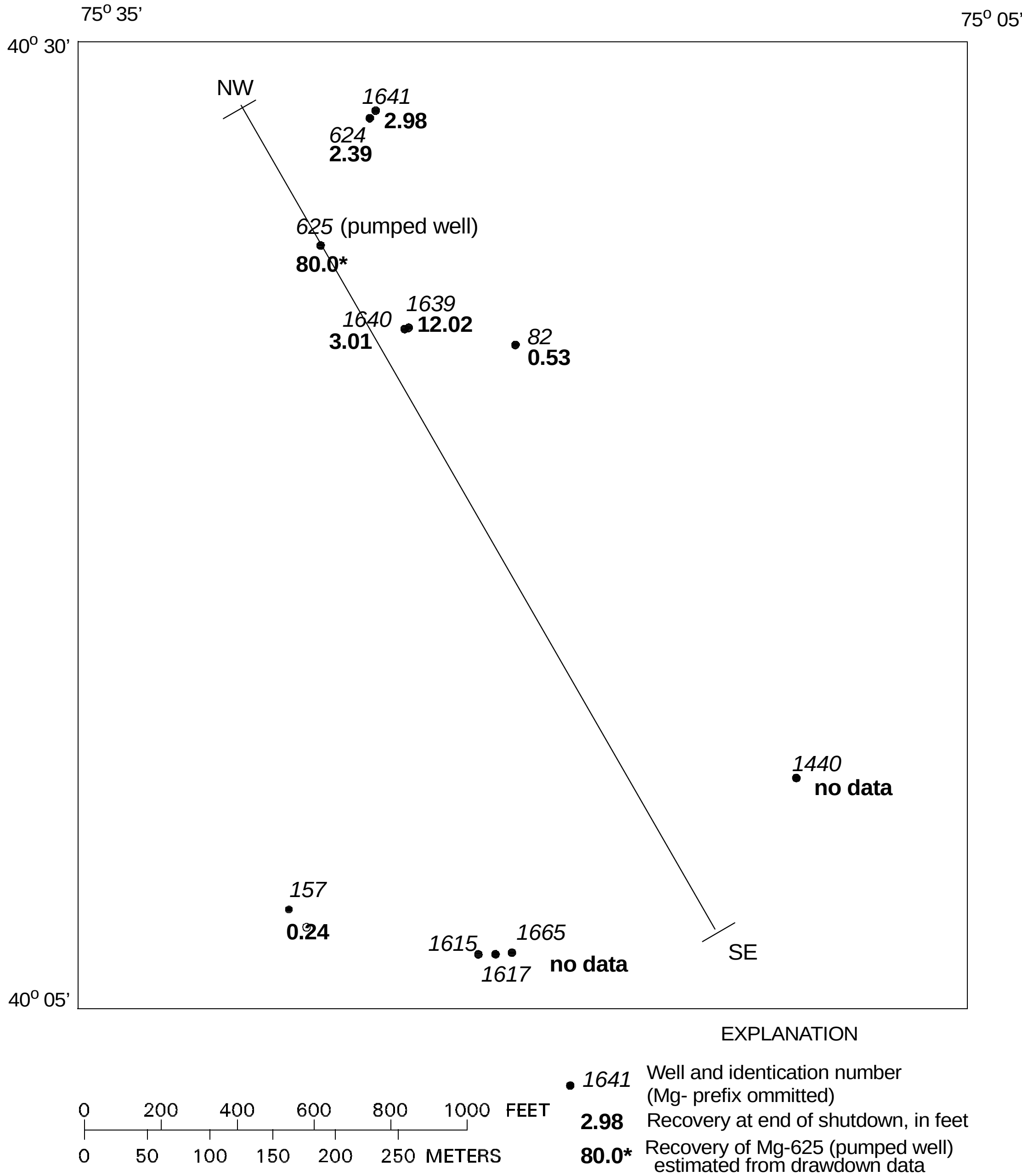


Figure 20.-- Well locations and recovery at end of shutdown of well Mg-625 at the J.W. Rex Co. property in northwestern Lansdale, Pa., October 27, 1997. Well Mg-625 normally pumped at a rate of 30 gallons per minute and was shutdown for 13 hours (water-level data from QST Environmental Inc., 1998).

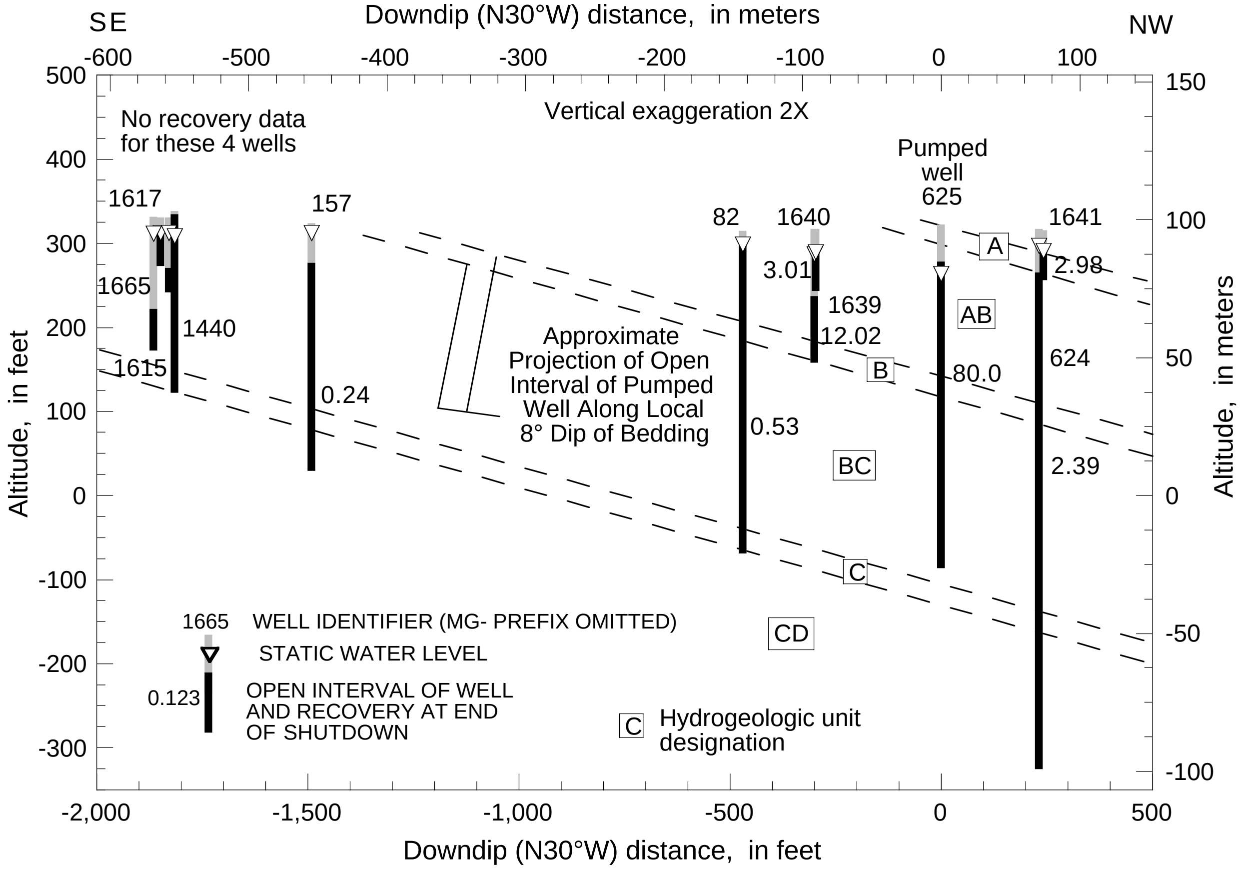


Figure 21.-- Cross-section of open intervals of wells, static water level, and recovery at end of shutdown at the J.W. Rex Co. property in northwestern Lansdale, Pa., October 27, 1997. Well Mg-625 normally pumped at a rate of 30 gallons per minute and was shutdown for 13 hours (water-level data from QST Environmental Inc., 1998). Also shown is the conceptual model of dipping high-permeability (A, B, C) and low-permeability (AB, BC, CD) beds. High and low-permeability beds stratigraphically above A are combined and designated "Bulk aquifer." All wells are projected onto a vertical plane parallel to the dip direction.

Water levels measured at Rex are influenced by pumping at well Mg-625 and at other nearby wells (fig. 22). Water levels recover most in well Mg-1639 and less in wells Mg-624 and Mg-1640 when Mg-625 is shutdown. The response of the water level in well Mg-1641 to pumping at Mg-625 is difficult to discern because of a greater response to pumping at a different well. This stronger influence is shown by the recovery of nearly 5 ft on October 25, 1997, while Mg-625 was pumping. The temporal pattern of this water-level change strongly indicates pumping effects and not recharge. Unfortunately, this other well was also apparently shutdown a few hours before Mg-625 on October 26, 1997. The magnitude of the water-level rise in Mg-1641 measured on October 26-27, 1997, is caused by shutdown of Mg-625 in addition to the other, unidentified well.

The shutdown of well Mg-625 is treated as negative pumping in the aquifer-test analysis. The water-level rise is computed as the water level after shutdown minus the water level at the time of shutdown. This approach assumes that the water levels are in equilibrium with pumping and other stresses and boundary conditions at the time of shutdown, and that the only factor causing water-level rise is the cessation of pumping. This assumption, while not perfectly met, is supported by the relatively constant water levels prior to shutdown on October 26, 1997 (fig. 22). The water-level rise for well Mg-1641 is computed by subtracting the water-level rise on October 25, 1997, during the apparent shutdown of a different pumping well, from the water-level rise on October 26-27, 1997, when both wells (Mg-625 and the unidentified well) are not pumping. These procedures yield the water-level rise and fall as a function of elapsed time since shutdown.

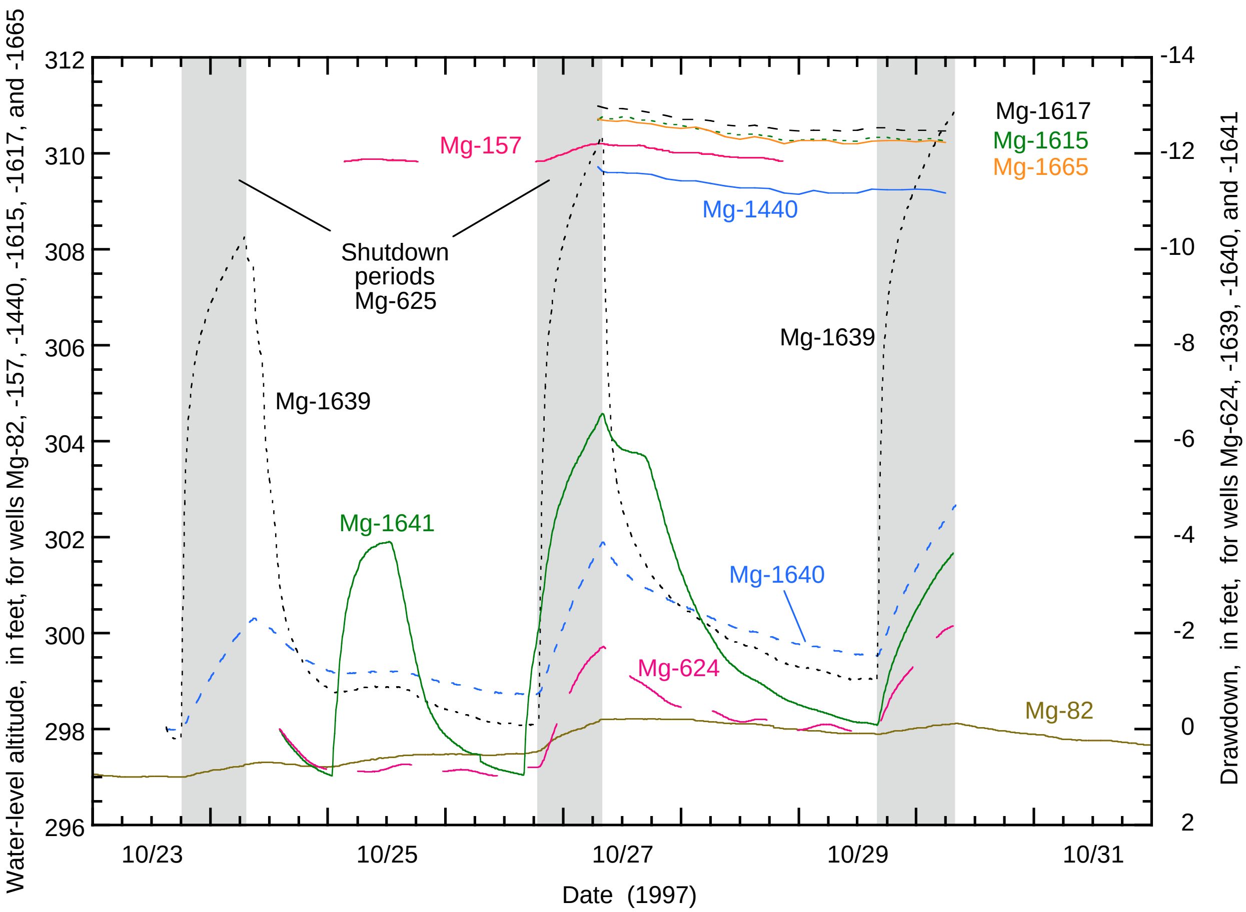


Figure 22.-- Measured water levels or drawdown in wells Mg-82, Mg-157, Mg-624, Mg-1440, Mg-1615, Mg-1617, Mg-1639, Mg-1640, Mg-1641, and Mg-1665 at the J.W. Rex Co. property in northwestern Lansdale, Pa., October 23-31, 1997. Well Mg-625 was pumped at a rate of 30 gallons per minute, except for the shutdown periods indicated (data from QST Environmental Inc., 1998).

Model Structure and Boundary Conditions

Local-scale models of ground-water flow in the northwestern part of Lansdale (fig. 23) are embedded within the regional-scale model. Most of the aquifer system of the regional-scale model is represented almost identically as done by Senior and Goode (1999). The top model layer represents soil and weathered bedrock and has uniform thickness of 19.7 ft (6 m). The bedrock aquifer is represented by two model layers, each 328 ft (100 m) thick. Most pumping and observation wells in the regional-scale model are located in the second model layer, as described by Senior and Goode (1999).

In the areas of Keystone and Rex, the horizontal grid spacing is reduced to allow simulation of dynamics of the local-scale flow system (see fig. 12 for example of grid spacing reduction near the pumped well). While the horizontal grid spacing is 328 ft (100 m) in the regional-scale model, the smallest horizontal spacing is 32.8 ft (10 m) at Keystone and 82 ft (25 m) at Rex. Because the local-scale models are embedded directly in the regional-scale model, regional fluxes are automatically simulated.

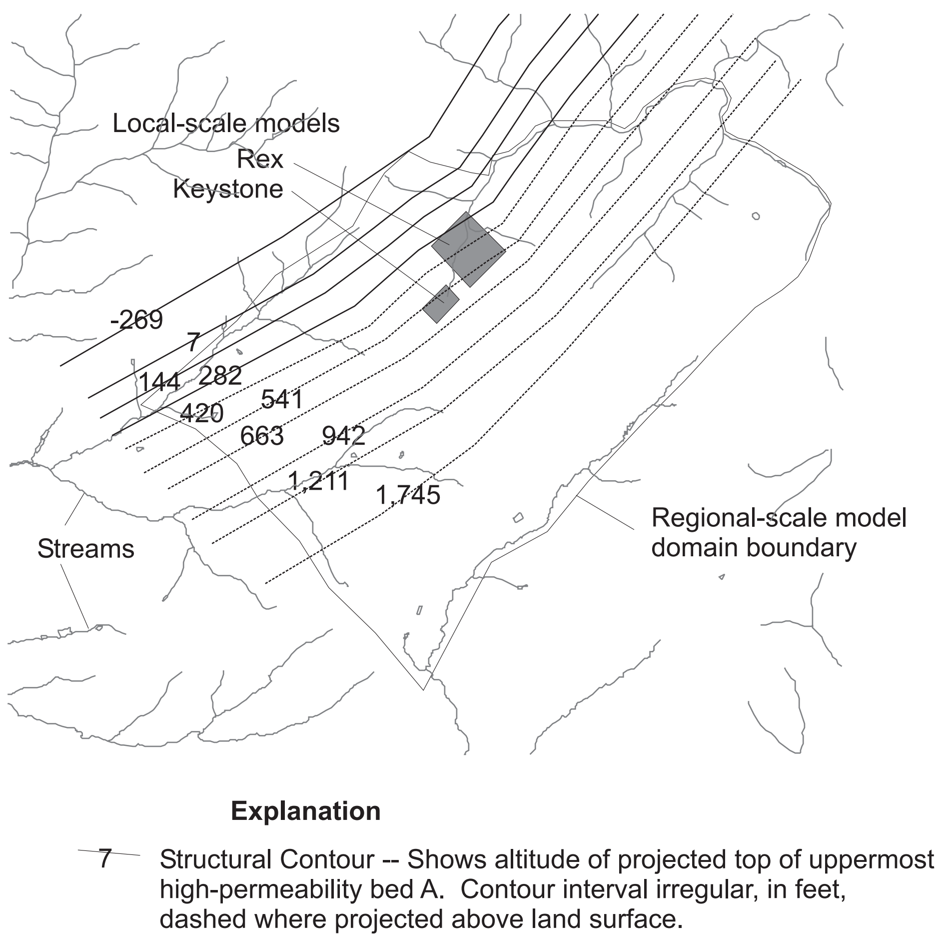
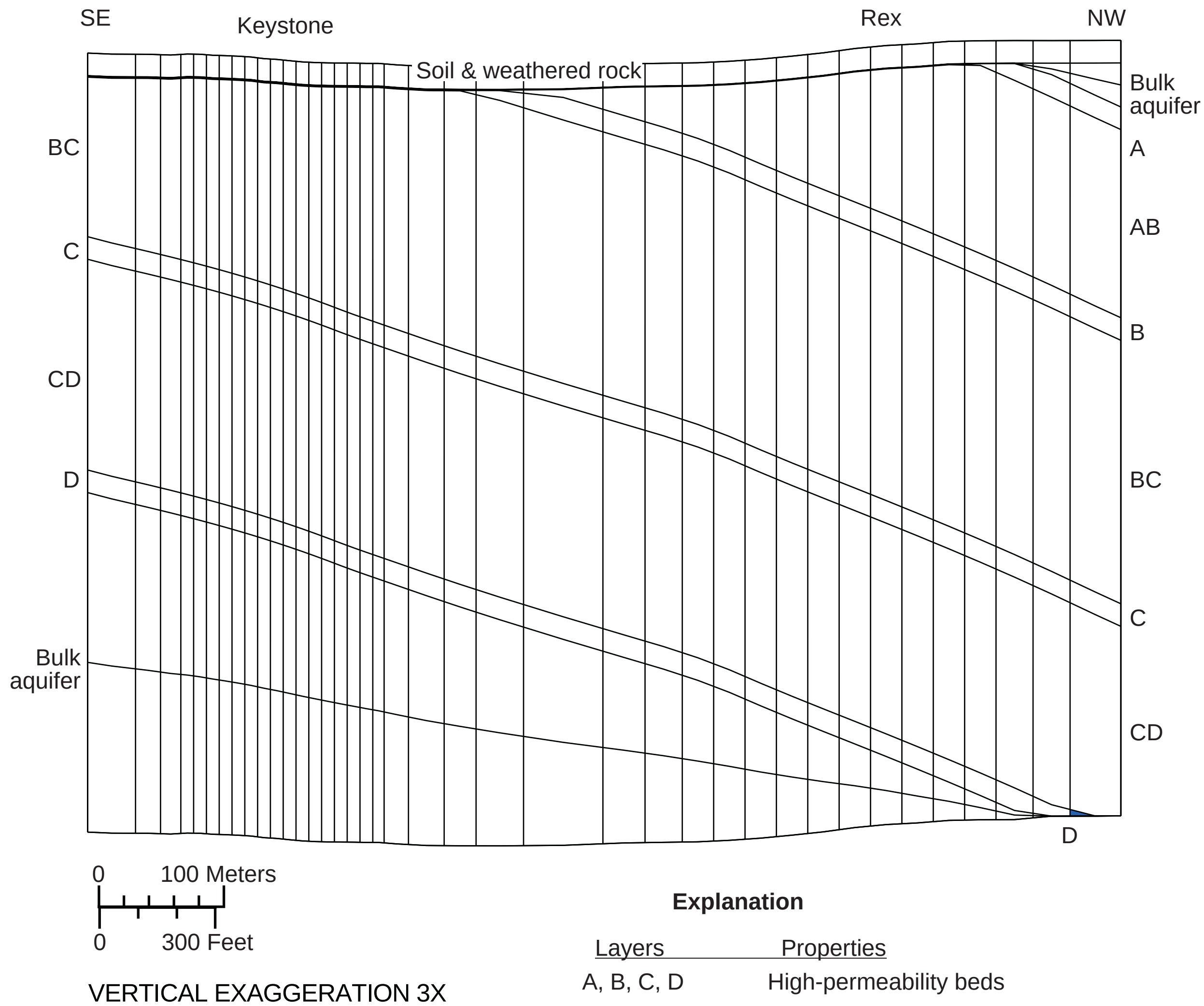


Figure 23.-- Location of the regional-scale model domain and the embedded local-scale models at the Keystone Hydraulics property (Keystone) and the J.W. Rex Co. property (Rex) in northwestern Lansdale, Pa. The contours are the elevation of the top of bed A, represented by model layer 4, taken from regional geologic maps and from local stratigraphic correlations of gamma logs (Conger, 1999).

The vertical discretization also is changed at Keystone and Rex to incorporate more detailed hydrogeologic information from aquifer-isolation (packer) and aquifer-test results. On the basis of these results and borehole-logging information, a conceptual hydrogeologic cross-section at Keystone and Rex is constructed (fig. 24). Not all geologic beds observed are treated separately in this model. Rather, the major beds that control the hydraulic connections between the pumping and observation wells at these sites are included. The strike and dip of beds is taken from regional information and from the local correlations (Conger, 1999) of gamma logs (fig. 23).



Explanation

Layers	Properties
A, B, C, D	High-permeability beds
AB, BC, CD	Low-permeability beds
Bulk aquifer	Undifferentiated Brunswick Group

Figure 24.-- Hydrogeologic cross-section in the downdip direction in northwestern Lansdale. Assumed high-permeability beds are designated A-D. Assumed structural contours of the top of bed A are shown in figure 23.

In summary, the model structure is similar to the regional-scale model of Senior and Goode (1999) but with finer horizontal spacing and heterogeneous structure in the northwestern Lansdale area. The bedrock properties are the same as the regional-scale model outside the area of the dipping beds in northwestern Lansdale. Where the dipping beds occur between the bottom of the soil / weathered rock layer and at a depth of less than 676 ft (206 m), the homogeneous bedrock is replaced by dipping beds of alternating high and low hydraulic conductivity. Computations associated with mapping the dipping bed structure into the three-dimensional model grid are handled by a preprocessor using a programmed GIS (geographic information system) (Shapiro and others, 1997; Winston, 1999).

Aquifer-Test Simulation

The local-scale model, which is embedded in the regional-scale model, is calibrated by simulation of aquifer-test results at Keystone and Rex. Drawdown during pumping at Keystone is simulated, similar to the approach of Senior and Goode (1999). In this case, however, the local-scale numerical model is used instead of the analytical model used by Senior and Goode (1999). Water-level rise during shutdown of well Mg-625 is simulated at Rex. The MODFLOWP calibration procedure is applied simultaneously for the two separate aquifer tests (table 4). This procedure yields estimates of model parameters for the entire model using information from both tests. The calibration results for the two locations are shown separately below, but only one model, with one set of optimum parameters, is used by both simulations.

Table 4.-- Optimum and approximate, individual, 95-percent confidence-interval values for hydraulic conductivity and specific storage for calibrated simulation of ground-water flow in northwestern Lansdale, Pa.

[ft / d, foot per day; ft^2 / d; feet squared per day]

			Approximate, individual, 95-percent confidence interval	
Parameter	Units	Optimum value	Lower value	Upper value
Beds A & B hydraulic conductivity	ft/d	7.2	5.7	9.2
Beds A & B transmissivity[1]	ft^2/d	142	112	181
Beds C & D hydraulic conductivity	ft/d	36	15	82
Beds C & D transmissivity[1]	ft^2/d	709	295	1,610
Interbeds AB & BC hydraulic conductivity	ft/d	9.6×10^{-4}	1.5×10^{-4}	5.4×10^{-3}
Interbed CD hydraulic conductivity	ft/d	.26	.087	.78
Beds A & B specific storage	per foot	1.2×10^{-6}	9.3×10^{-7}	1.5×10^{-6}
Beds C & D specific storage	per foot	1.4×10^{-6}	1.4×10^{-7}	1.3×10^{-5}
Interbeds AB & BC specific storage	per foot	9.9×10^{-8}	1.8×10^{-8}	5.6×10^{-7}
Interbed CD specific storage	per foot	4.2×10^{-6}	6.4×10^{-7}	2.8×10^{-5}

[1] Transmissivity is product of hydraulic conductivity and bed thickness [19.7 ft (6 m)].

The estimated hydraulic parameters from this three-dimensional numerical model can be qualitatively compared to parameters identified previously using analytical models. The estimated transmissivity of each isotropic high-permeability bed at Keystone is 709 ft^2/d, and the sum of the transmissivities is about 1,420 ft^2/d. Senior and Goode (1999) used a single-layer anisotropic analytical model and estimated the equivalent isotropic transmissivity as 2,300 ft^2/d, about 60 percent higher. QST Environmental, Inc. (1998) estimated single-layer isotropic transmissivity at Rex of 160 to 665 ft^2/d. Goode and Senior (1998) estimated transmissivity as 320 ft^2/d from recovery of Mg-625 at Rex. The sum of the transmissivities of high-permeability beds A and B estimated here is about 280 ft^2/d.

The calibrated model can approximately simulate measured drawdown during the aquifer test at the Keystone site (fig. 25). Compared to the analytical model match using the anisotropic model by Senior and Goode (1999), this model does not match the measured drawdowns as well. The shape and relative magnitudes of drawdown at the four wells used in the analytical analysis (fig. 25 A) are reasonably matched, although the absolute drawdowns are off by up to 20 percent. The drawdown at the end of the test is matched for Mg-1619, but the simulated drawdown curve is steeper than that measured. Well Mg-1611, which is immediately adjacent to the pumped well, but shallow, is not matched. In the model, this well is located in a low-permeability interbed and is poorly connected to the pumped interval. The measured drawdown suggests a higher-permeability connection between this well and the pumped interval, which is not included in the model. However, if this well is moved downward to the pumped bed, drawdown would be higher than that measured. This match could be improved by adding more local heterogeneity to the model.

The calibrated model can also approximately simulate measured water-level recovery during the shutdown test at Rex (fig. 26). The match is best for Mg-1639, which recovered the most. The differences between the measured and simulated recovery are most evident for wells that have smaller recovery. The shapes and relative magnitude of recovery in Mg-624 and Mg-1641 are matched well, but the absolute magnitudes are too low in the model. The slope of the recovery curves at Rex are steeper than the drawdown curves at Keystone, indicating a difference in hydraulic properties at the two locations.

The hydraulic conductivities estimated by local-scale model calibration in the northwestern Lansdale area include alternating beds of high and low hydraulic conductivity (table 4). Water levels change the most in wells open to high-permeability beds that also intersect the pumped well. Water levels generally do not change as much in wells that are open to beds above or below the pumped bed, in particular if the well is isolated by more than one intervening low-permeability bed. The storage coefficients are generally low, reflecting the confined conditions in the bedrock aquifer.

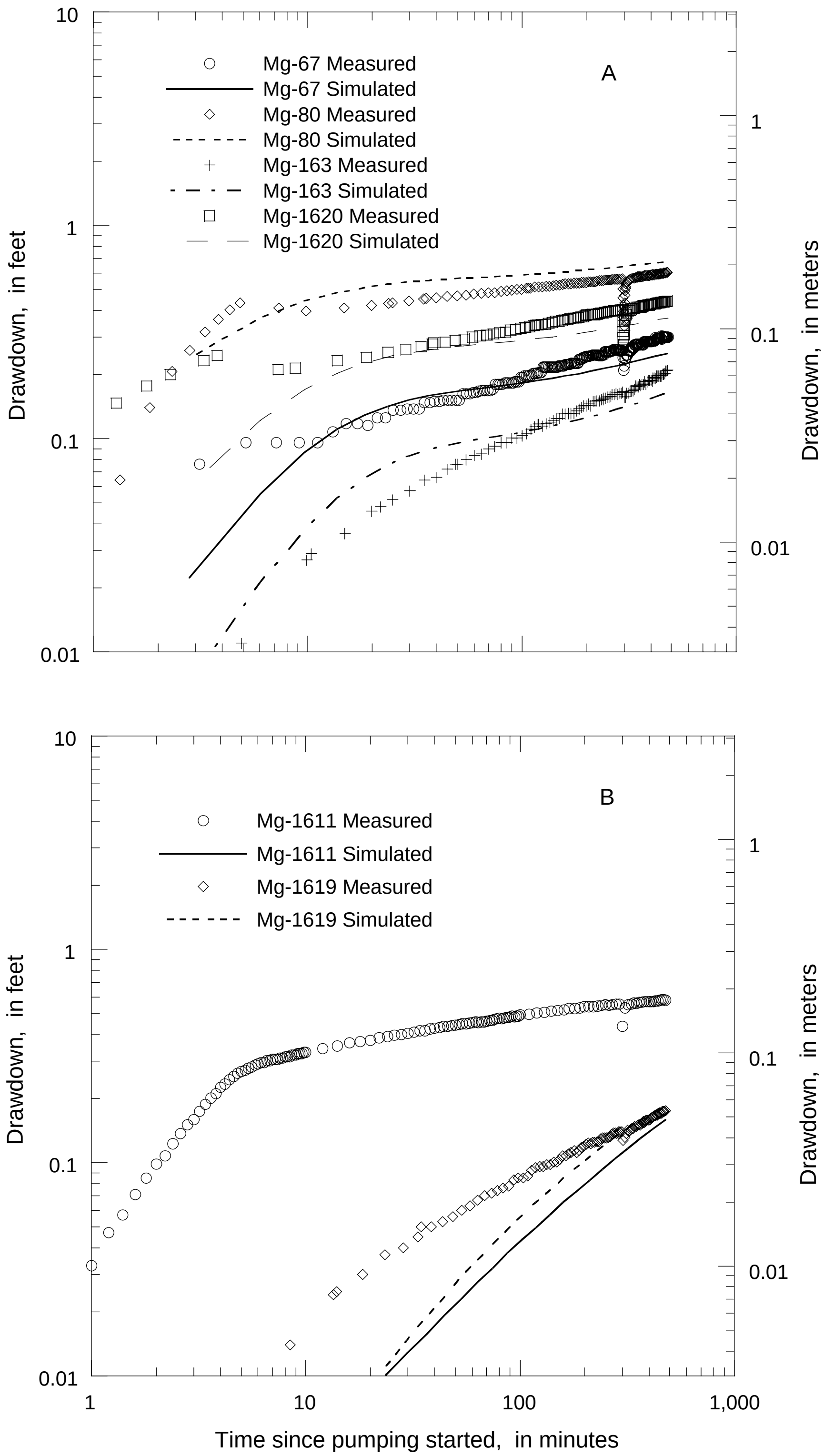


Figure 25.-- Measured and simulated drawdown in wells (A) Mg-67, Mg-80, Mg-163, Mg-1620, and (B) Mg-1611 and Mg-1619 at the Keystone Hydraulic property in northwestern Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours.

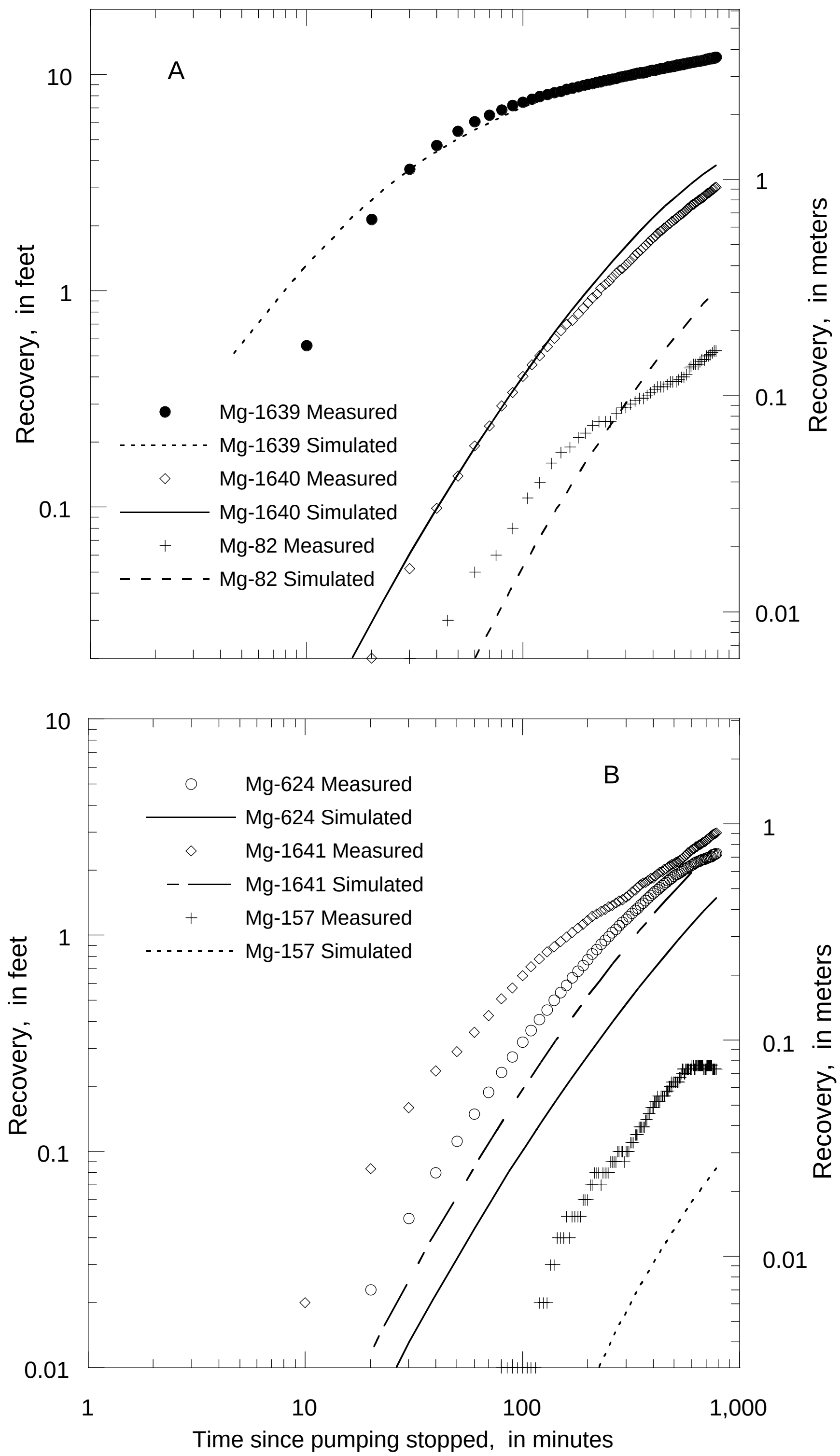


Figure 26.-- Measured and simulated water-level recovery in wells (A) Mg-82, Mg-1639, Mg-1640, and (B) Mg-157, Mg-624, and Mg-1641 at the J.W. Rex Co. property in northwestern Lansdale, Pa., October 26-27, 1997. Well Mg-625 was shutdown after pumping for 60 hours at a rate of 30 gallons per minute.

Effect of Pumping on Ground-Water Flowpaths

Steady-state-flow fields are simulated to determine the effects of pumping on ground-water flowpaths. The first scenario is for 1997 pumping conditions, which include continuous pumping in northwestern Lansdale at the J.W. Rex Co. production well. The second and third scenarios include additional pumping at a rate of 10 gal/min (gallons per minute) at wells Mg-1610 and Mg-1620, respectively. In areas where bedding is explicitly represented by the model, well pumpage is evenly distributed among high-permeability beds (table 5) intersected by the well. The pumped layers for wells Mg-625, Mg-1610, and Mg-1620 are the same as used in the analysis of the aquifer-test results.

Table 5.-- Annual average pumping rates for wells in and near Lansdale, Pa. during 1997.
[USGS, U.S. Geological Survey; --, not numbered]

USGS local well number Mg-	Owner	Owner well number	Model cells			Pumping rate (gal/min)
			Layer(s)	Row	Column	
625	J.W. Rex Co.	1	6	17	64	30.0
1610	Keystone Hydraulics	1-I	8	33	43	[1]10.0
1620	Keystone Hydraulics	2-S	8	44	46	[2]10.0
498	North Penn Water Authority	L-23	2,4,6,8	11	32	25.0
593	North Penn Water Authority	L-25	6,8,10,11	17	29	34.1
69	North Penn Water Authority	L-10	8,10,11	52	27	68.1
914	North Penn Water Authority	NP-12	10,11	55	101	54.9
566	Lehigh Valley Dairy	5	11	57	19	64.4
59	Lehigh Valley Dairy	3	11	57	19	44.4
1418	Ziegler	--	10,11	57	80	4.4
140	Lehigh Valley Dairy	4	11	59	17	92.5
1125	North Penn Water Authority	NP-61	11	60	77	71.5
875	North Wales Water Authority	NW-17	11	62	88	71.0
1051	North Wales Water Authority	NW-22	11	62	94	136.3
1198	Merck & Co.	PW9	11	65	5	26.1
125	Merck & Co.	PW2	11	71	3	[3]94.1
130	Merck & Co.	PW7	11	73	10	91.0
171	Precision Tube	1	11	73	23	6.4
204	Precision Tube	2	11	73	24	6.4
126	Merck & Co.	PW3	11	75	6	96.7
77	North Penn Water Authority	L-18	11	76	36	67.2
75	North Penn Water Authority	L-16	11	76	72	43.5
124	Merck & Co.	PW1	11	76	3	48.4
202	North Penn Water Authority	L-22	11	76	27	37.6
76	North Penn Water Authority	L-17	11	77	29	40.4
73	North Penn Water Authority	L-14	11	77	55	38.5
78	North Penn Water Authority	L-19	11	77	62	31.9

[1] Pumping included only for scenario “Mg-1610”.
[2] Pumping included only for scenario “Mg-1620”.
[3] Pumping rate at cell is (rate at PW2) + [(rate at PW8) / 2],

The hydraulic conductivity for the upper two interbeds is adjusted for the flowpath simulations. Simulations with the calibrated bed parameters (table 4) yield water levels too high where interbeds AB and BC are present in the model. These beds have very low hydraulic conductivity in the calibrated model, reflecting conditions at Rex. Use of these values, however, for interbeds that extend throughout the entire regional model causes a hydraulic barrier to flow that is inconsistent with observed water levels. For the flowpath simulations here, the hydraulic conductivity of interbeds is increased by two orders of magnitude. This hydraulic-conductivity parameter has the largest relative confidence interval, indicating that it is the least constrained by the measurements. Conceptually, this increase reflects the limited extent of low-hydraulic-conductivity parts of the interbed, or localized fracturing of the interbeds. With this increase, these units still have the lowest hydraulic conductivity in the model.

Pumping well Mg-625 under 1997 Conditions

The simulated contributing area of well Mg-625 is an area of about 3,000,000 ft^2 (fig. 27). Recharge at the well and in an area south of the well is simulated as flowing into the well. A thin region of capture also occurs to the northwest and north and extends more than 3,300 ft from the pumped well. These results reflect the location of the well near the ground-water divide between recharge flowing northeast into tributaries of the Neshaminy Creek and recharge flowing southwest towards tributaries of the Towamencin Creek. These results suggest that pumping at Mg-625 is effective in capturing water infiltrating at Rex south of the pumping well, but that infiltration falling on parts of the property to the north may not be captured by Mg-625 (see fig. 4 for locations). Recharge located downgradient from the pumped well, within about 150 ft of the well, is not captured by the well.

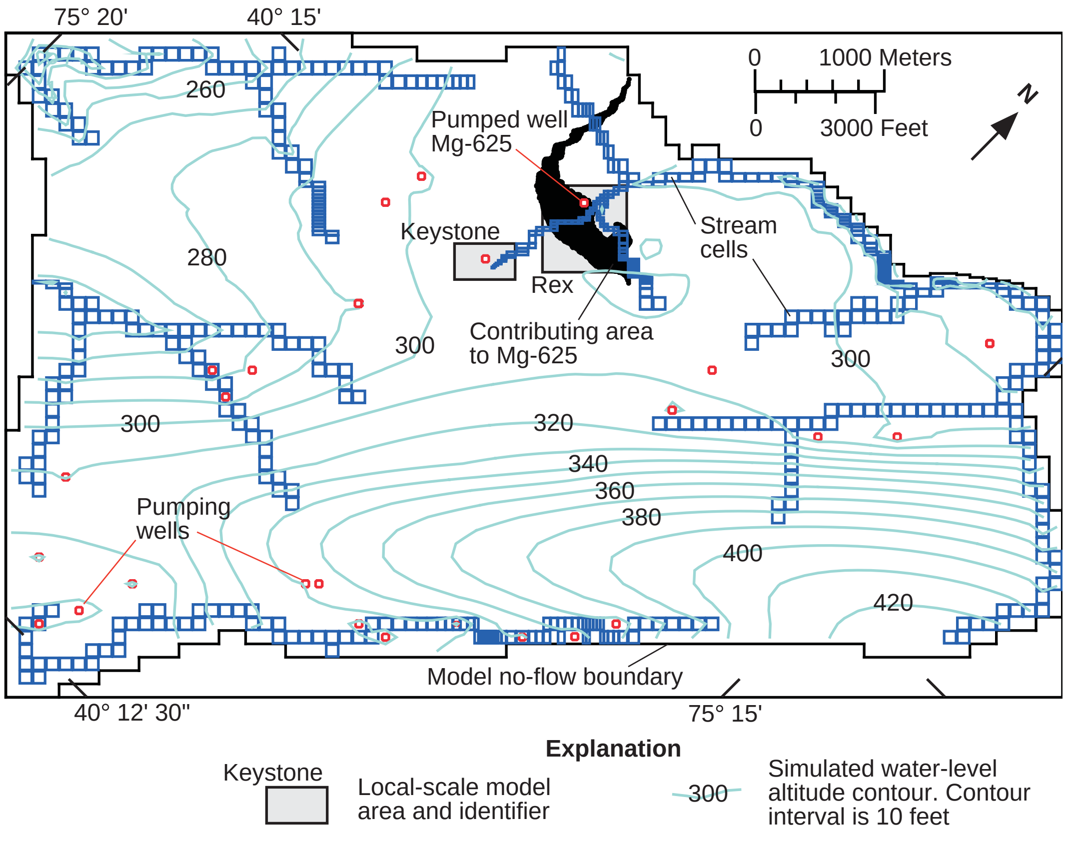


Figure 27.-- Simulated water levels in the upper part of the bedrock aquifer in Lansdale, Pa,. and vicinity for 1997 conditions and contributing area for well Mg-625 pumping at a rate of 30 gallons per minute.

The shape and location of the contributing area is determined by regional flow patterns and the geologic structure of beds of high and low hydraulic conductivity in the vicinity of the pumping well. To illustrate the role of geologic structure, the contributing area simulated with the model developed here is compared to the contributing area simulated using the regional-scale model of Senior and Goode (1999). The size of the contributing areas is identical because the same pumping rate and recharge rate are used. The regional-scale model includes anisotropy to approximate the effects of dipping beds but does not include individual high- or low-permeability beds. Compared to the contributing area simulated with the local-scale model, the regional-scale model contributing area has a larger extent in the strike direction, the direction of preferred flow in the anisotropic model (fig. 28). In the regional-scale model, which used a horizontal grid cell size of 328 ft (100 m), recharge on the surface within about 600 ft of the pumping well is captured. In contrast, the local-scale model, with finer resolution and explicit dipping beds, indicates that recharge as close as a few tens of feet in the downgradient direction (north) will not contribute to the well. The contributing areas to alternative pumping wells, located more to the north or in shallow layers, also could be simulated using this model to compare the effectiveness of capturing recharge in areas that do not contribute to Mg-625.

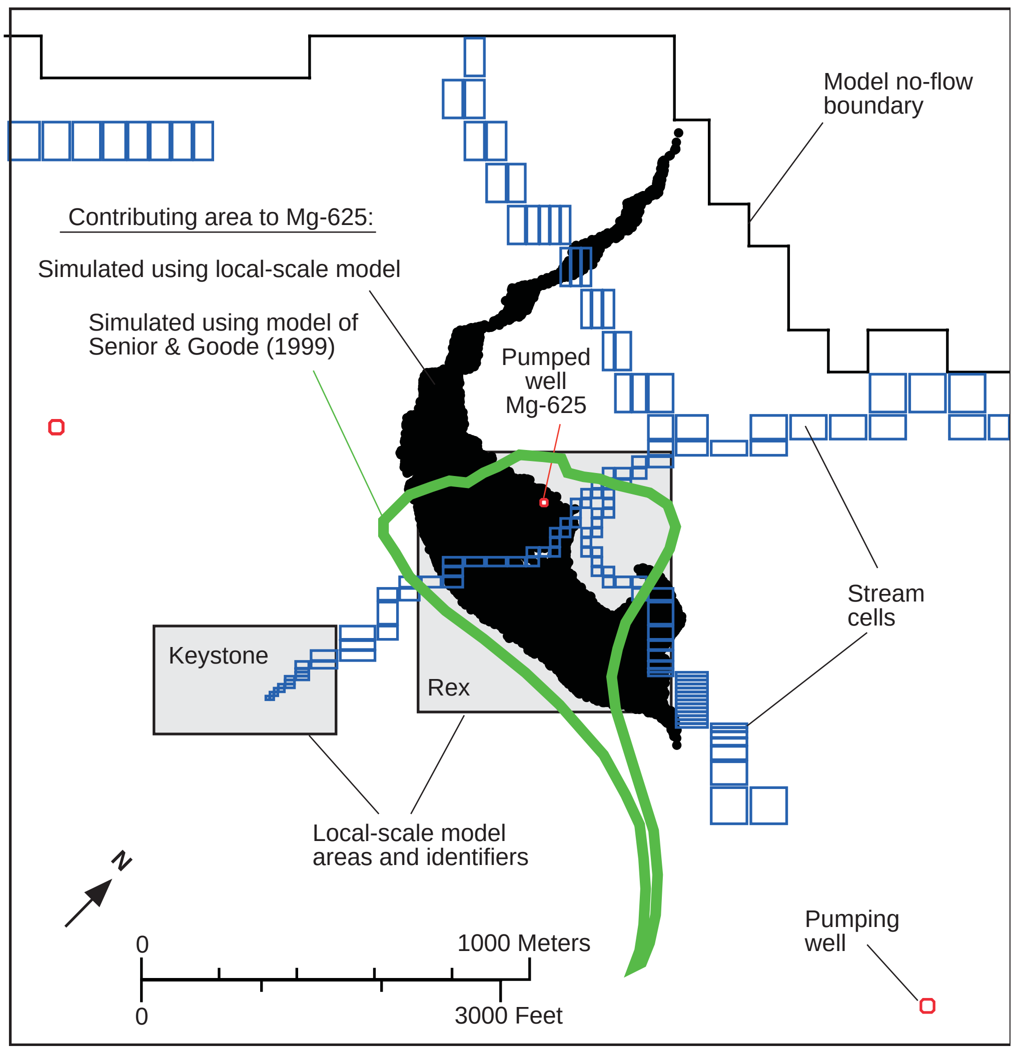


Figure 28.-- Contributing areas for well Mg-625 pumping at a rate of 30 gallons per minute in Lansdale, Pa., for 1997 conditions simulated using the local-scale model developed here and the regional-scale model of Senior and Goode (1999). See figures 27 and 4 for locations.

Additional pumping at well Mg-1610

The simulated contributing area of well Mg-1610 is an area of about 1,000,000 ft^2 (fig. 29). The contributing area includes the land surface at the well location and extends about 1,500 ft east in the upgradient direction. This pumping well is located southwest of the simulated ground-water divide, hence the contributing area is somewhat simpler than that for well Mg-625. These results suggest that infiltrating water northeast and east of the pumping well at the Keystone property would contribute to Mg-1610.

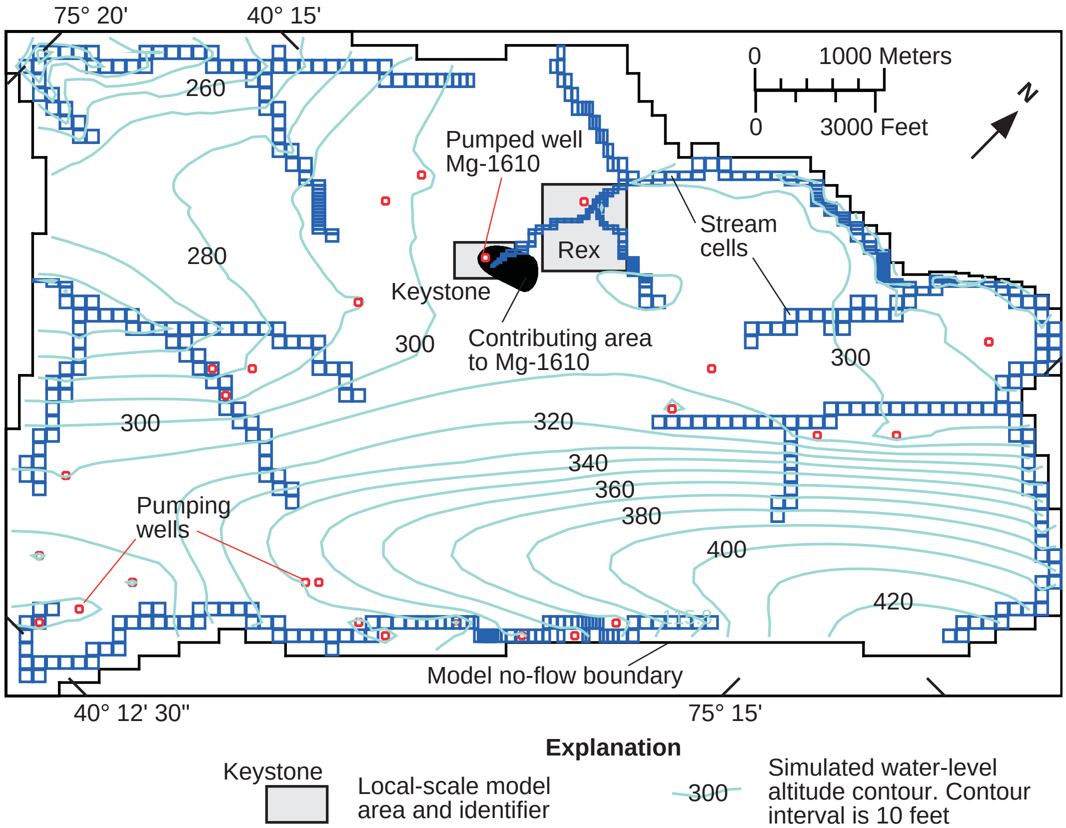


Figure 29.-- Simulated water levels in the upper part of the bedrock aquifer in Lansdale, Pa,. and vicinity, and contributing area for well Mg-1610 pumping at a rate of 10 gallons per minute.

Additional pumping at well Mg-1620

The simulated contributing area of well Mg-1620 is an area of about 1,000,000 ft^2 (fig. 30). The contributing area is very similar to that for well Mg-1610, which is deeper than Mg-1620, but which is open to the same high-permeability bed. The contributing area for Mg-1620 is shifted to the east and south, relative to the contributing area for Mg-1610.

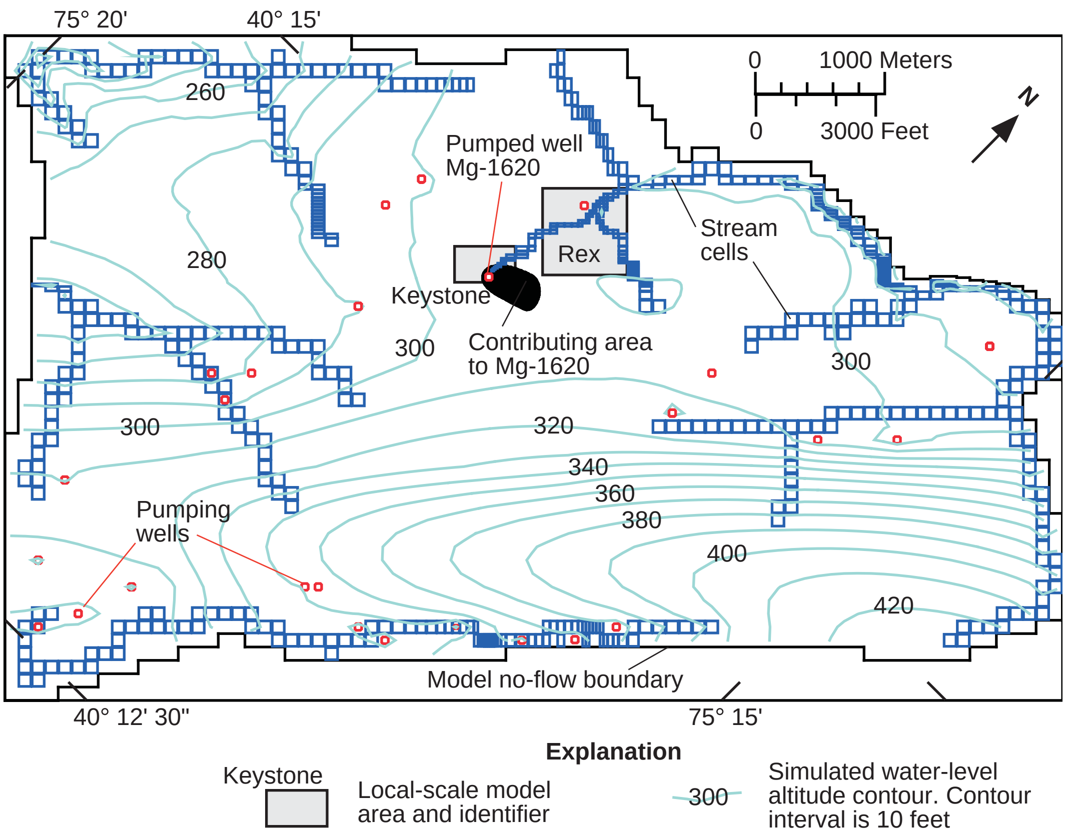


Figure 30.-- Simulated water levels in the upper part of the bedrock aquifer in Lansdale, Pa., and vicinity, and contributing area for well Mg-1620 pumping at a rate of 10 gallons per minute.

SUMMARY AND CONCLUSIONS

Ground water in the Lansdale area is contaminated with organic solvents, and contaminant migration is of concern as pumping patterns in the area change. The U.S. Geological Survey (USGS) provided technical assistance from 1995 to 1999 to the U.S. Environmental Protection Agency (USEPA) in remedial investigations at the North Penn Area 6 site, Lansdale, Pa. This assistance included describing the ground-water system and simulating of ground-water flow on a regional scale. Lansdale is underlain by dipping beds of Triassic-age shales, siltstones, and sandstones of the Brunswick Group and Lockatong Formation that form a layered aquifer. The calibrated regional model indicated that the aquifer appeared to be anisotropic; transmissivity is greatest in the strike direction of underlying rocks. However, the regional model did not incorporate local hydrogeologic detail that could be inferred from geophysical logs, aquifer tests, and other data. In late 1999, the USGS began additional modeling of ground-water flow in selected areas of Lansdale on a local scale.

On a local scale, ground-water flow in the Lansdale area occurs primarily in high-permeability bed-oriented features that dip northwest with the regional geologic structure. Numerical models of local-scale ground-water flow are developed using the dipping hydrogeologic structure and measured water levels during aquifer tests at three locations in north-central and northwestern Lansdale. Predictive contributing areas for individual pumping wells are simulated using the calibrated local-scale flow models, in conjunction with a previously developed calibrated steady-state model of regional flow.

The local-scale flow models approximately simulate measured drawdown and recovery during pumping. The heterogeneous hydrogeologic structure is explicitly incorporated in the local-scale models, and the hydraulic properties of the dipping beds are identified by model calibration. The complex three-dimensional structure in the models cannot be included in simpler analytical models commonly used for aquifer-test analysis. Transmissivities determined from calibration of the local-scale models, nevertheless, are similar to those calculated using analytical methods for the aquifer tests. Computations associated with mapping the dipping bed structure into the three-dimensional model grid are handled by a preprocessor using a programmed geographic information system. Reduced flow across low-permeability beds is explicitly simulated. The dipping high-permeability beds are extensive in the strike direction but are of limited extent in the dip direction. This model structure yields ground-water-flow patterns characteristic of anisotropic aquifers; preferred flow is in the strike direction. The transmissivities of high-permeability zones in the local-scale models range from 142 to 1,900 ft^2/d (13 to 177 m^2/d). The hydraulic conductivities of low-permeability parts of the aquifer range from 9.6×10^{-4} to 0.26 ft/d (2.9×10^{-4} to 0.079 m/d). In comparison, the geometric mean or “effective” transmissivity of the regional scale model was 1,050 ft^2/d, a value that lies in the range of transmissivities determined from the local-scale model and represents bulk properties of the Brunswick Group aquifer. Storage coefficients and specific-storage values are very low, indicating the confined nature of the aquifer system.

Predictive contributing areas indicate the general characteristics of ground-water flow towards wells in the Lansdale area. Recharge to wells in Lansdale generally comes from infiltration near the well and over an area that extends upgradient from the well. Locally, all recharge in the area of the pumping well is not captured; recharge in the downgradient direction about 150 ft from the pumping well will flow to other discharge locations. The contributing areas for two wells pumping at 10 gal/min (Mg-1609 in north-central and Mg-1610 in northwestern Lansdale) extend about 1,500 ft upgradient from the wells. The contributing area is more complex at ground-water divides and can extend in more than one direction and capture recharge from more than 3,300 ft away, for pumping at a rate of 30 gal/min (well Mg-625 in northwestern Lansdale). The contributing area for well Mg-625 in northwestern Lansdale is compared for simulations using the local-scale model and the previously developed regional scale-model. The contributing area simulated by the regional-scale model is more elongated in the strike direction than the contributing area simulated by the local-scale model. Differences in the shape of the contributing area simulated by the two models may be caused partly by grid-size differences and the anisotropic character of the aquifers in the regional-scale model.

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